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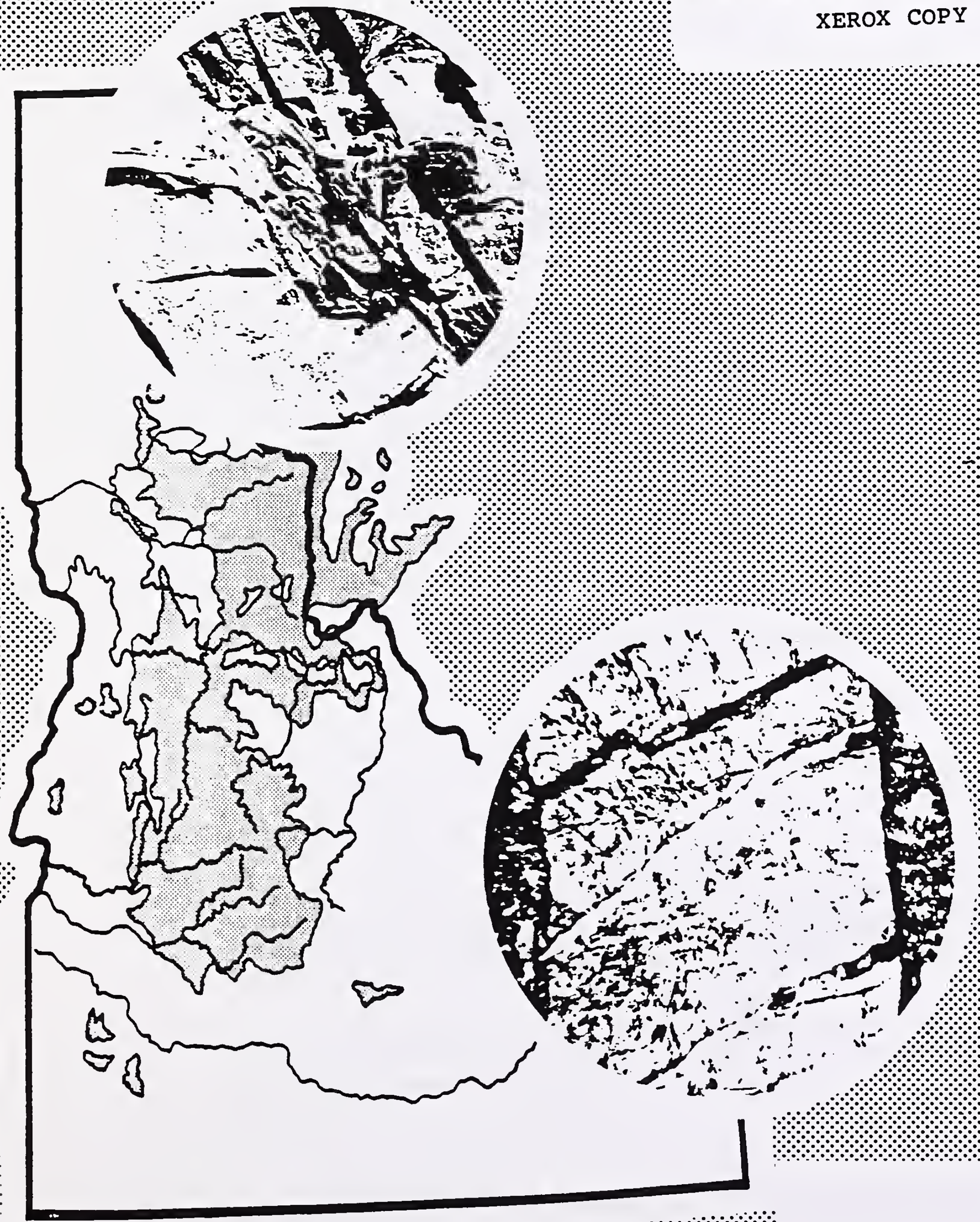

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SOIL AND BEDROCK PROPERTIES: WEATHERING AND ALTERATION PRODUCTS AND PROCESSES IN THE IDAHO BATHOLITH

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RESEARCH SUMMARY

The mechanical properties of granitic rock and the derived weathered materials are intimately related to slope stability in the Idaho batholith. Processes and products of physical weathering, chemical weathering, and hydrothermal alteration of batholithic rocks can be ranked in terms of degree of weathering and used to predict slope stability.

Bulk density, hardness, unconfined compressive strength, and velocity of sonic wave transmission of weathered rock samples are all strongly correlated with weathering properties. Following subtle, initial hydrolysis of biotites, physical weathering processes are dominant in the breakdown of rock to grus at and near the soil surface. Grus is often formed with little alteration of biotites and feldspars to secondary minerals or amorphous oxides of silicon, aluminum, and iron.

Bedrock buried at depths below the rooting zone is subject to intense chemical weathering and mineral alteration. The degree to which chemical weathering processes can proceed is related to the inherent slope stability of the site. On gentle slopes with low erosion rates, chemical weathering can advance to a point where the weathered product consists of quartz grains embedded in a matrix of clays and crystalline and amorphous oxides of silicon, aluminum, and iron.

X-ray diffraction patterns of powdered bedrock samples show decreases in peak intensity, diffuseness of d-spacings, and complete absence of some reflections with increasing weathering. Intermediate stages of chemical weathering of individual minerals can result in clay minerals that are not in equilibrium with current pedogenic conditions favoring formation of kaolinite, halloysite, and illite. We have found iron-rich smectite-iddingsite alteration after biotite and interstratified clays of undetermined origin.

Low temperature hydrothermal alteration is common in the Idaho batholith, although it is frequently difficult to determine on the basis of mineralogy alone. The frequency of occurrence of hydrothermal alteration is associated with major and minor structural lineaments mapped by satellite imagery. Alteration products are often located in clay seams bounded by shear zones. Mineralogies may be dominated by smectites and zeolites (clinoptilolite), but often the clays may be kaolinite, halloysite, and minor mixed-layer illite-montmorillonite. Altered zones with this latter more normal group of clay minerals may reflect a combination of surface weathering and hydrothermal alteration.

The morphology of Idaho batholith soils is not particularly suited for predictions of subsurface weathering characteristics. The strong topographic influence on slope stability, coupled with youthful soils resulting from high erosion rates, masks or obliterates the soil morphology-bedrock weathering relationship. The relationship between the pH of the B and C horizons and the degree of bedrock weathering is fairly well defined: the pH decreases from 6.3 to 4.9 as the parent rock becomes increasingly weathered. Other parameters, such as clay, free silica, and sesquioxide content, and soil color were not strongly correlated with weathering.

CONTENTS

NOTE: A Weathering Guide for field use containing information from the section BEDROCK WEATHERING is appended inside the back cover.

INTRODUCTION

Batholiths are composite masses of granitic rocks that have areas ranging from tens to thousands of square kilometers. Batholiths generally cut sharply across their wall rocks and are surrounded by contact metamorphic aureoles clearly demonstrating formation by intrusion of magma from greater depths than the surrounding rock. Contacts between plutons and the adjoining country rocks are vertical or steeply dipping over distances measured in thousands of meters. The contacts are commonly gradational due to mutual exchange of material between intruded granite and country rock. In places, the granite at the boundaries passes into a zone of migmatite, consisting of metamorphosed country rock veined and streaked with intruded granite or pegmatite (Turner and Verhoogen 1960). The main characteristics of batholiths are enormous size, method of emplacement, discordant relationship to the country rock, and lack of a visible floor.

The Idaho batholith is a typical batholith (fig. 1). It outcrops across Idaho more or less continuously for about 250 miles (400 km) north-south and 80 miles (130 km) east-west (fig. 2). Larsen and others (1954) have dated the intrusion at 90 to 100 million years BP[1], indicating emplacement during the early and middle Cretaceous. The batholith was later uplifted by faulting and exposed by erosion, forming the present group of mountain ranges in central Idaho.

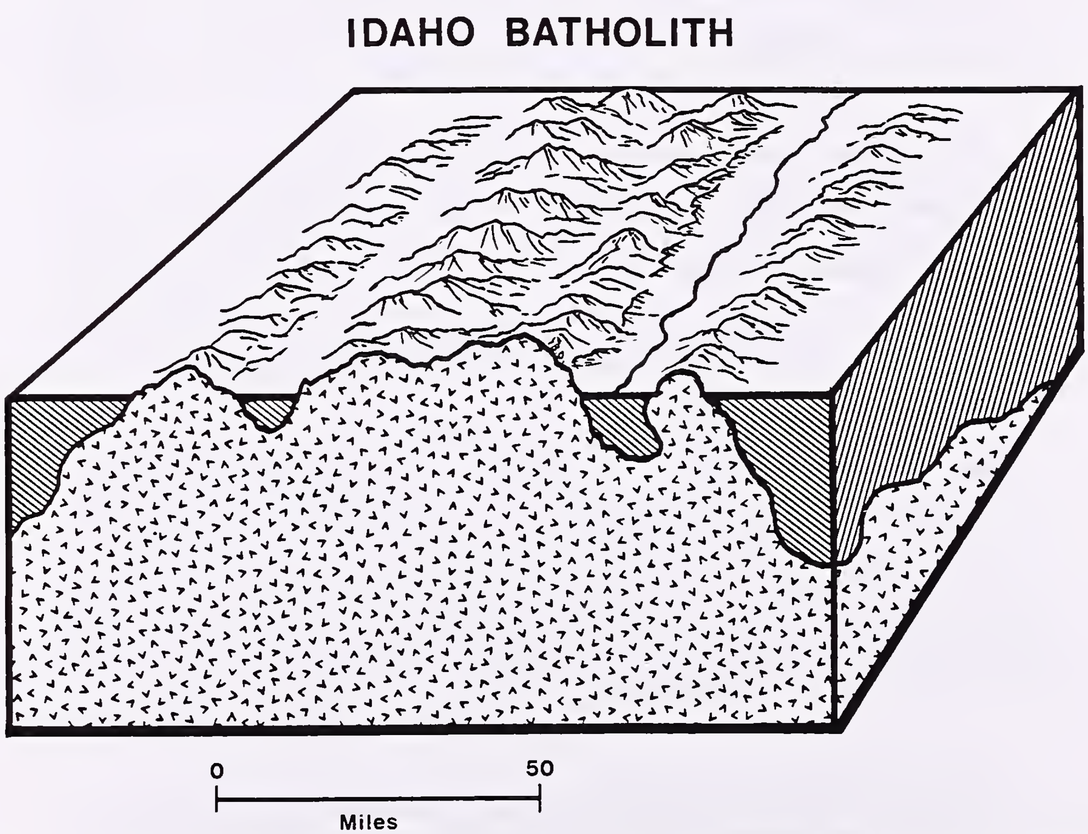

Figure 1.--Typical bedrock relationship showing the discordant relationship between the Idaho Batholith and the country rock. Uplift subsequent to emplacement was followed by the removal of thousands of feet of overburden. (Vertical scale is exaggerated.)

[1]Years before present, used to express age of geologic or soil deposits.

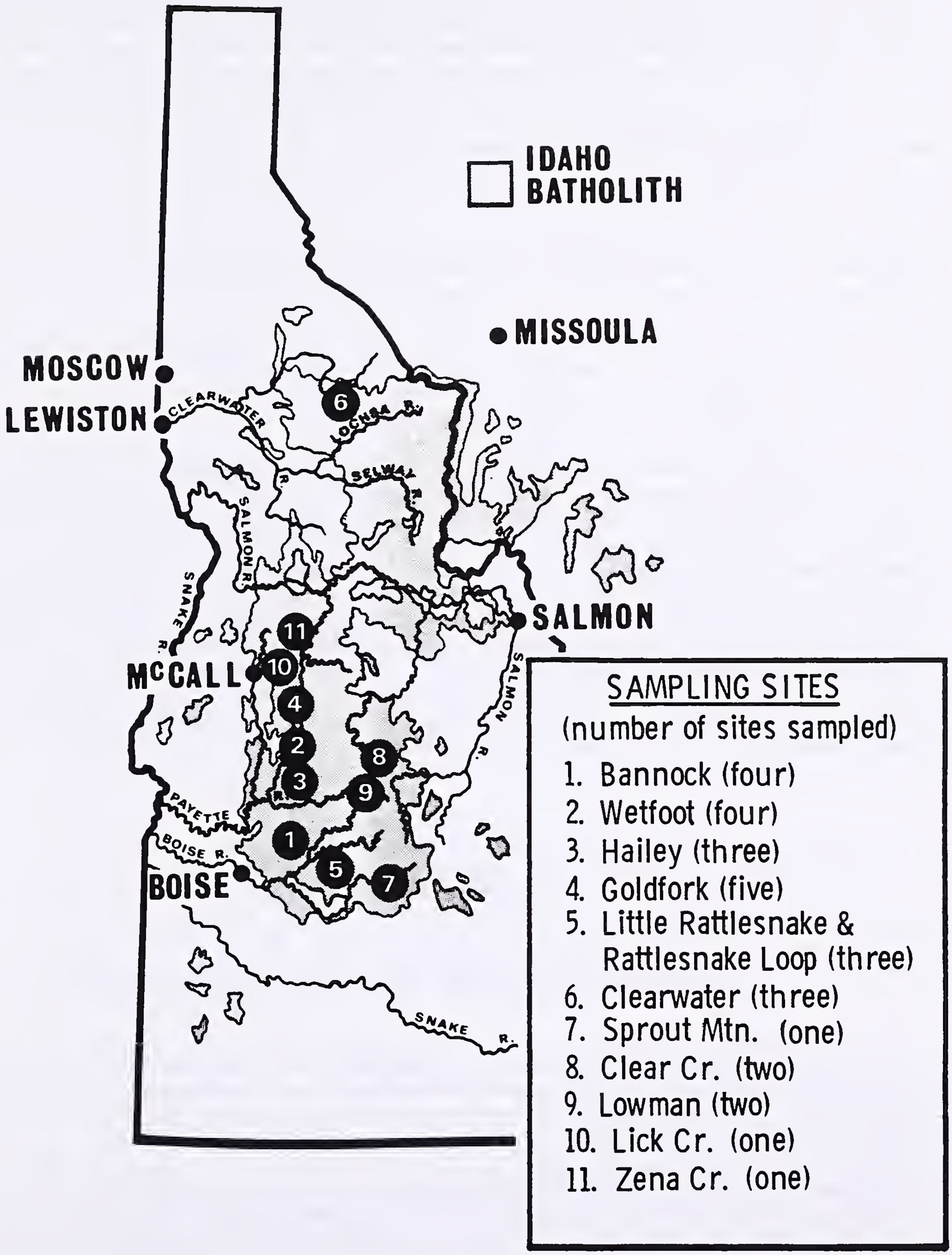

Figure 2.--Map of Idaho showing the location of the Idaho batholith and sampling sites within the batholith.

Much of the batholith is overlain by forest land administered by the Forest Service. Timber, water, forage, wildlife, and recreation are resources that support a substantial portion of the economic and social welfare of Idaho and its surrounding States. At present, the optimum utilization of these resources cannot be realized because of man's damaging activities, principally those related to logging and road construction. When imposed on the unstable landscapes of the batholith these activities have the potential of causing tremendous increases in erosion and sedimentation (Megahan and Kidd 1972a, 1972b; Rice and others 1972).

Streams of the Idaho batholith contain a sizeable resident game fish population and provide spawning and rearing grounds for anadromous fish; however, anadromous fish populations generally have been on the decline since about 1957 as a result of dam construction along the Columbia and Snake Rivers and increased sediment loads in rivers draining the Idaho batholith (Ortman 1967; Platts 1972; Mallet 1974).

Slope instability and resulting sedimentation have been matters of increasing concern to Forest Service land managers in the Northern and Intermountain Regions over the last two decades. Today, these problems are of paramount importance in decisions affecting multiple use of National Forest lands.

The relations between rock and soil properties and erosion in the batholith have been recognized as important, but largely unknown quantities in land management. In 1970, the Intermountain Forest and Range Experiment Station funded a 3-year cooperative venture between Howard University and the Station's research laboratory in Boise, Idaho. The purpose of the study was to investigate the engineering, mineralogical, chemical, and hydrological properties of soil and bedrock in the Idaho batholith. This paper deals primarily with the chemical and mineralogical properties of rocks, weathering products, and soil, and sheds light on rock weathering processes and soil genesis. Relationships between weathering products, and engineering and hydrologic properties of soil and bedrock are discussed in less detail. Some conclusions about the relationships between weathering and hydrothermal alteration processes and slope stability are also discussed.

BACKGROUND

Description of the Idaho Batholith — Geologic Setting

The Idaho batholith lies mainly in central Idaho, but extends in a northeasterly direction into Montana. Its surface outcroppings cover an area of about 16,000 mi^2 (41 000 km^2). The batholith is a member of a chain of large intrusive bodies extending along the western border of North America.

Much of the Idaho batholith belongs to an inner facies of rather uniform, light-colored granitic rock, mainly of granodiorite composition (Ross 1963). An outer mass or border zone of more calcic rock may have originally constituted an envelope that enclosed much of the main facies (Ross 1963). The Idaho batholith is zoned spatially and chemically; older, more mafic rocks are found at the western edge, becoming progressively more felsic and younger in a easterly direction (Bennett 1974)[2].

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[2]Bennett, Earl H., II. 1974. The general geology of that part of the Northern Rocky Mountain province containing the Idaho batholith. Unpubl. Rep. USDA For. Serv., Intermt. Reg., 228 p. Ogden, Utah.

Schmidt (1964) defined the zoning sequence from west to east on a transect near Cascade, Idaho. He distinguished a migmatite border zone, a quartz diorite gneiss zone, a leucocratic quartz diorite zone, a granodiorite zone, and, finally, a quartz monzonite zone, some 40 miles (65 km) east of the migmatite border. The batholith intrudes sedimentary and volcanic rocks ranging in age from Precambrian to Triassic.

A modal analysis of rock types in the batholith presented by Ross (1963) indicates that samples from the main inner facies are mainly granodiorite, although many are quartz monzonite. These analyses indicate that the main mass of the batholith may be somewhat more calcic than was thought previously (for example, see Larsen and Schmidt 1958).

Elevations in the batholith range from about 2,000 ft (730 m) where the Salmon River leaves the batholith to more than 10,000 ft (3 650 m) in the south-central part of the batholith. Because of the wide range in elevation, numerous climates are encountered in the batholith. Geomorphic features reflect both the tectonic setting and climatic diversity of the area.

The tectonic history of the Idaho batholith and adjoining areas has been important in controlling the gross landforms of the area. The Snake River downwarp and parallel trend of compensatory uplift north of the river plain control the major drainage systems within the entire State. Large-scale faulting associated with these tectonic events, and stream piracy virtually reversed the direction of flow of the Salmon River and turned its waters from the Missouri system to the Snake and Columbia Rivers (Anderson 1947). North-south striking faults along the western Idaho border tend to control valley development and drainage patterns in the western batholith. The northwest trending faults and structural control in general in eastern Idaho have controlled the orientation of drainages and of several mountain ranges, including the Bitterroot Range in the northeastern batholith.

Alpine glaciation played a major role in shaping the landscape during Quaternary times. Landforms in the Sawtooth Mountains, Stanley Basin area, the Salmon River Range east of McCall, Idaho, and the Bitterroot Mountains have all been modified to varying degrees by alpine glaciation.

In contrast, areas of lower elevation have landforms showing the influence of fluvial slope-forming processes. Deeply incised canyons and steep valley side slopes epitomize the landscapes. The area along the Middle Fork of the Salmon River in the central batholith provides a good example of such landscapes. Here, the combined effects of a major stream carrying a considerable sediment load and a changing base level associated with mountain uplift and the Snake River downwarp have produced the deeply incised canyons of the Salmon River Breaks area.

Eolian deposits are common in many areas of the northern Idaho batholith, and traces have been reported in the interior of the batholith as far south as the head-waters of the Deadwood River, east of Cascade, Idaho. The source of much of the loess in the northern batholith is glaciofluviatile material from the Columbia River Basin that was reworked by winds during the late Quaternary. Some loess is also of volcanic origin. Tephra falls from both Glacier Peak (12,000y BP) and Mazama (6,600y BP) have been reported in the northern Idaho batholith (Wilcox 1965), and tephra deposits are present in the south-central batholith east of Cascade, Idaho (Clayton and Wendt 1975), but their origin and age remain obscure.

Climatic Setting

Two basic precipitation sources operate within the region of the Idaho batholith. Most winter precipitation originates from cyclonic storms that develop over the Pacific Ocean. These storms travel on an easterly course over Washington and Oregon, losing much of their water content over the mountain barriers of the Cascade Range. Thus a distinct rainshadow effect reduces the precipitation delivered to the Idaho batholith region. Both the frequency and the intensity of winter cyclonic storms increase from south to north (Dean 1974)[3], and the rainshadow effect is less pronounced north of the Clearwater River.

Summer high-altitude convective storms developed from moist air moving north from the Gulf of Mexico are common in the region. These storms are of short duration, but precipitation intensities usually exceed those developed from cyclonic storms. Storm intensities of 3 inches/hour (7.6 cm/hour) have a probable return period (p = 0.5) of 4 years in the central Idaho batholith (Kidd 1964). The frequency of summer convective storms decreases from south to north (see footnote 3 on this page).

Mean annual precipitation in the batholith ranges from below 15 inches (38 cm) to over 70 inches (180 cm). The driest portions are the southern fringe of the batholith, the South Fork of the Boise River, and the main Salmon River Canyon in the central batholith. Mean annual precipitation in excess of 50 inches (127 cm) is rare in the southern half of the batholith, the only extensive area located in the Sawtooth Mountains. Mean annual precipitation in extensive areas of the batholith north of the Clearwater River exceeds 50 inches (127 cm) (see footnote 3).

Air temperatures within the Idaho batholith are most strongly influenced by topographic variations. Average annual temperatures decrease approximately 4.1° F (2.3° C) per 1,000-ft (365-m) gain in elevation over the range 3,600 ft (1 000 m) to 9,850 ft (3 000 m) above sea level. Mean monthly temperatures at all stations are highest in July and lowest in January. Average monthly maximum temperatures range from a high of 92° F (33° C) in July at several stations in the southern batholith below 4,000 ft (1 460 m) in elevation to a low of 26° F (-3° C) in January at Cobalt, elevation of 6,810 ft (2 480 m). The lowest average monthly minimum temperatures range from 0° F (-18° C) in January at Chilly Barton Flat, elevation 6,175 ft (2 250 m), to 33° F (1° C) in July at Warren, elevation 5,907 ft (2 154 m).

METHOD

Sample Sites and Field Procedures

Sampling sites for this study were selected to insure geographical, elevational, and climatic variability. Preliminary observations suggested that various landforms have distinctive bedrock weathering characteristics. For example, landscapes glaciated in late Pleistocene or Neoglacial time tend to be underlain by relatively hard, unweathered bedrock. In contrast, more highly weathered bedrock is commonly associated with landscapes that have undergone fluvial dissection and associated slope processes less erosive than glaciation. Sites were selected to provide a representative array of weathering and fracture density classes as described by Clayton and Arnold (1972).

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[3]Dean, E. Nelson. 1974. Some climatic and hydrologic characteristics data for physiographic sections of that portion of the Northern Rocky Mountain Province containing the Idaho batholith. Unpubl. Rep. USDA For. Serv. Intermt. Reg., 141 p. Ogden, Utah.

Fifteen sites were selected for sampling and testing in 1970 and 14 in 1971. The sites were located in four National Forests: Boise, Clearwater, Payette, and Sawtooth. Location of the sampling areas and the number of sites sampled at each area are shown in figure 2.

We initially tried to sample bedrock by core drilling, but abandoned this procedure because of poor recovery of highly weathered and (or) fractured bedrock. The only acceptable alternative to core drilling was to select and sample sites on fresh roadcut faces. A fresh roadcut was defined as one not exposed more than 2 months prior to sampling.

Rocks were hand-sampled at a variety of depths below the original surface and measured vertically from the top of the cut face. During 1970 sampling, several rock samples were taken systematically at predetermined depths from each face. Because of difficulty in assigning a single value of weathering (Clayton and Arnold 1972) at the sites selected in 1970, site selection and sampling methods were changed in 1971. Roadcut faces were then chosen for uniformity of weathering and fracturing along the entire exposed face, and a random sample of several rocks representative of the exposed face were collected and treated compositely as a single sample representative of the site.

At each site, a soil pit was dug above the roadcut and generally 15 to 60 feet (5 to 20 m) from the face. This soil is presumed to be formed from the parent material exposed in the roadcut. A description of the soil in the walls of each pit included the following: horizons, depth, color, structure, presence of clay films, percent by volume of stones and rocks, roots, pores, and horizon boundaries. Textures and pH were determined in the laboratory.

Site variables described included location, field classification of parent rock (lithology, weathering, and fracture density classes), surface stone and rock outcrop percent, landform, slope, aspect, elevation, and a description of the type and amount of erosion. We described the trees, shrubs, forbs, and grasses present and estimated relative amounts of vegetation (percent by number of stems for trees and shrubs) for the site as a whole. We assigned the vegetation to habitat types using the classification system proposed by Steele and others (1975).

Duplicate core samples were taken from each soil horizon by driving a brass cylinder (136.4 cc volume) horizontally into the side of the exposed soil pit. Horizons that contained too many stones damaged the sampling apparatus and therefore were not sampled for bulk density. In addition, grab samples were collected from all major horizons for laboratory analyses.

Laboratory Procedures — Soil

All soil samples were air-dried, then sieved through a 2 mm diameter sieve. Most analyses were performed on the <2 mm fraction of the soil. Water retention at 15 bars was determined with a pressure membrane apparatus. Retentions at 1/3 and 0.1 bar tension were determined with a porous plate pressure chamber device. Percent sand, silt, and clay are expressed as percentages by weight of the <2 mm fraction. They were determined by the hydrometer method in a sedimentation cylinder. The 136.4 cc soil core samples were ovendried and weighed to the nearest 0.01 of a gram in order to determine bulk density. Soil pH values were determined on the <2 mm fraction on a 1:1 soil paste. Values reported are averages of duplicate samples. Cation exchange capacities were determined in duplicate on <2 mm soil samples. This analysis employed saturation of the exchange complex with sodium followed by removal of the sodium by means of

ammonium acetate. Sodium was determined by flame emission spectroscopy. Free iron (Fe_2O_3) was determined on soil samples (<2 mm) after removal of organic matter with 30 percent H_2O_2. A citrate-dithionite extraction was employed and iron determined colorimetrically with o-phenanthroline. Free silica (SiO_2) was determined on the clay fraction (<2 µm) of soils. Organic matter was removed by digestion in 30 percent H_2O_2. Clay samples were then deferrated using a citrate-dithionite extraction after buffering to pH 7.2 with $NaHCO_3$. Samples of deferrated clays were dried and free SiO_2 was removed by boiling for 5 minutes in a 2 percent Na_2CO_3 solution. Silica was determined by the molybdate blue method. Free alumina (Al_2O_3) was determined on the same extract for a few samples. Aluminum was determined by atomic absorption spectroscopy utilizing a N_2O oxidant.

Organic matter determinations were run on paired samples in a carbon combustion furnace (900° C) using continous oxygen flow.

Clays (<2 µm) were separated from the <2-mm fraction by sedimentation, then sucked onto ceramic plates for mounting in the diffractometer. This method produces samples with basal cleavage orientation. Copper Kα radiation was used to produce the x-ray diffractograms. A variety of heat treatments and ethylene glycol solvation were used to confirm presence of the various clay minerals. Peaks on the diffractograms were marked and, for each mineral assemblage, the relative intensities were semiquantitatively indicated. Minerals were considered present either in trace amounts, minor amounts, abundant amounts, or dominant amounts.

The sand and silt fraction left after decanting the clay for x-ray analysis was sieved. The heavy minerals were then separated from the fine sand fraction (0.25 to 0.10 mm) with bromoform and mounted on microscope slides. These slides were examined for etching of the mineral grains and mineral alteration.

Laboratory Procedures — Rock

Bulk density determinations were taken on each rock sample prior to subsampling by weighing, coating with paraffin, and measuring the volume by water displacement.

Rock samples that had been pulverized in a ball mill for 5 minutes were analyzed for free Fe_2O_3. The powdered rock was analyzed the same way that soil samples were, except that predigestion in H_2O_2 was omitted.

Descriptions of each sample included: (1) presence of iron oxide staining and source mineral, (2) color of biotites and appearance of biotite grain boundaries, (3) degree of opacity of feldspars, (4) frequency of intergranular fractures, and (5) extent of argillation by color, and degree of staining of mineral faces with a 1 percent solution of p-aminophenol (method after Dodd 1955).

Thin sections were prepared from rocks from all sites. Two hundred point counts were made to provide a petrologic description for each sample. Sections were examined for individual mineral weathering products and degree of weathering, but no point counts were made to quantify the amount of weathering. Instead, a visual estimate of percent weathering of primary mineral types was made. We also described mineral pleochroism, twinning (relict in secondary minerals), grain structure and orientation, and undulose extinction if present (principally in quartz).

Rock samples were pulverized in a ball mill, sieved to pass a 60-mesh (d = 0.25 mm) screen and sprinkled on vaseline-coated slides. These slides were x-rayed using copper Kα radiation scanning the range 3° 2Θ through 32° 2Θ. Although this method was of little value in identifying weathering products, it did provide information about degree of weathering of primary feldspars.

Data from all field descriptions and laboratory procedures are reported by Hampton and others (1974a). X-ray diffraction studies on the soils are described by Clayton (1974). Selected soil and rock properties related to engineering research, results of seismic studies and hydraulic conductivity testing of fractured bedrock are presented by Hampton and others (1974b).

BEDROCK WEATHERING

The transition from fresh bedrock to weathered bedrock involves numerous chemical and physical processes. These processes manifest themselves in the physical and mineralogical (chemical) properties of the rock itself.

Granitic rock such as that found in the Idaho batholith is formed under tremendous pressures (1000 to 1500 bars) and high temperatures (600 to 800 °C). Rock formed under these conditions is in a state of disequilibrium when brought to the surface.

The overriding driving force for weathering processes is a continued readjustment the rock must make toward a thermodynamic equilibrium dictated by the changing environment the bedrock is subjected to when brought to the earth's surface. As Ollier (1969) points out, it is not necessary to assume that cooled magma at depth ever achieved a true thermodynamic equilibrium; only that this rock, when brought to the earth's surface, is less in equilibrium with surface conditions than its potential weathering products.

The physical properties and mineralogical changes reflecting various stages or degrees of weathering are numerous. To satisfy one intent of this study, to relate bedrock weathering to slope stability, we spent considerable effort in relating degree of rock weathering to mechanical strength and other engineering properties (Hampton and others 1974a; Hampton and others 1978).

We classified the samples in the field on the basis of semiquantitative degree of weathering (weathering class) and have used this as the dependent variable when field and laboratory measurments are compared. The seven weathering classes of Clayton and Arnold (1972) were used because they give reasonble field estimates of rock strength and secondary mineral formation. Many of the secondary mineral assemblages occur as a result of hydrothermal alteration or a combination of hydrothermal alteration and weathering. These processes will be distinguished whenever possible. The seven classes are described as follows:

Class 1. Unweathered Rock.--Unweathered rock will ring from a hammer blow; cannot be dug by the point of a rock hammer; joint sets are the only visible fractures; no iron stains emanate from biotites; joint sets are distinct and angular; biotites are black and compact; feldspars appear to be clear and fresh.

Class 2. Very Weakly Weathered Rock.--Very weakly weathered rock is similar to Class 1, except for visible iron stains that emanate from biotite; biotites may also appear to be "expanded" when viewed through a hand lens; feldspars may show some opacity; joint sets are distinct and angular.

Class 3. Weakly Weathered Rock.--Weakly weathered rock gives a dull ring from a hammer blow; can be broken with moderate difficulty into hand-sized rocks by a hammer; feldspars are opaque and milky; no root penetration; joint sets are subangular.

Class 4. Moderately Weathered Rock.--Moderately weathered rock may be weakly spalling. Except for the spall rind, if present, rock cannot be broken by hand; no ring or dull ring from hammer blow; feldspars are opaque and milky; biotites usually have a golden yellow sheen; joint sets are indistinct and rounded to subangular.

Class 5. Moderately Well Weathered Rock.--Moderately well weathered rock will break into small fragments or sheets under moderate pressure from bare hands; usually spalling; root penetration is limited to fractures, unlike class 6 rock where roots penetrate the rock matrix; joint sets are weakly visible and rounded; feldspars are powdery; biotites have a light-golden sheen.

Class 6. Well-weathered Rock.--Well-weathered rock can be broken by hand into sand-sized particles (grus); usually, it is so weathered that it is difficult to determine whether or not the rock is spalling; roots can penetrate between grains; only major joints are preserved and filled with grus; feldspars are powdery; biotites may appear as thin silver or white flakes.

Class 7. Very Well Weathered Rock.--Very well weathered rock has feldspars that have weathered to clay minerals; rock is plastic when wet; no resistance to roots.

The initial process of unloading overburden material during uplift and erosion is extremely important to all subsequent weathering processes. Small stress fractures and joints extant in the cooled magma at depth are expanded during overburden removal (Ollier 1965). These fractures provide the necessary pathways for water, the key ingredient in both physical and chemical weathering, both of which are generally near-surface processes.

The distinctions between physical and chemical weathering in the Idaho batholith are sometimes obscure. They are contemporaneous processes and complement each other synergistically. A typical example of this is provided by biotite weathering.

Initial weathering of individual biotite grains involves hydrolysis of nonframework potassium, and hydrolysis and oxidation of iron and magnesium. Removal of iron and magnesium oxides ensues (Walker 1949); the biotite grain forms fringed edges, expands and allows increased water entry. The increased water entry results in greater strain on the mineral because of hydration of weathering products and freeze-thaw mechanisms. This strain allows entry of water, weak organic acid solutions, and chelating agents that result in framework destruction and subsequent secondary mineral formation. Expanded biotite grains can exert enough pressure on the surrounding rock to fracture it, or at least to weaken bonds between adjacent minerals, and thus start the transition from rock to grus. Framework destruction and subsequent expansion is the principal process in the conversion from weathering class 1 to class 4, and will be described in more detail later in this section.

Physical Weathering

The documentation of physical weathering as a singular process, distinct from chemical weathering, requires recognition of (a) physical breakdown of the rock (loss of strength) and (b) a lack of secondary mineral formation. These two conditions were never met in our rock weathering research. Microbrecciated rocks from shear zones frequently appear in hand specimen to be physically, but not chemically, weathered rocks. Such shear zones are common throughout the Idaho batholith and can easily be recognized by their abrupt contact with boundary rock of greater competence. In thin section, however, rock from shear zones is readily recognized by the degree of microbrecciation, marked undular extinction of quartz grains (strain deformation) and, often, presence of alteration products, such as epidote and garnets.

Physical weathering probably is the major cause of physical weakening in rocks from weathering class 2 to class 4 or 5. The processes that cause this loss of strength are all surface or near-surface phenomena, with freeze-thaw cycles and hydration-dehydration being the most important processes. However, freeze-thaw cycles probably are inconsequential below a soil depth of 12 inches (30 mm) and intergranular stresses resulting from hydration and dehydration probably are restricted to the top 6 to 10 feet (2 or 3 m) of bedrock, except along the faces of joint systems.

Chemical weathering requires that water be present, but there is no requirement for periodic drying. Chemical weathering is therefore favored in subsurface zones that are periodically wetted by deep seepage, but are not subject to evapotranspirational water loss.

Chemical weathering does aid in the physical weathering of some rocks. This has been observed by engineers and those concerned with road maintenance, who contend that granite bedrock weathers, decomposes, or "air slakes" within a year or two after construction. Hence, rock that required blasting for road construction becomes a sediment source and maintenance problem. Such rock has weathered chemically by hydrolysis and, possibly, by oxidation (Henin and Pedro 1965) of biotites and initial hydrolysis of feldspars, but it has not undergone physical weathering until it is exposed in a roadcut.

The chemical weathering at depth, which may only be visible as mild iron oxide staining and slight opacity of feldspars, has resulted in a network of fine intergranular fracturing. Thus, the rock is primed for water entry and will lose its strength rapidly upon surface exposure. Rock samples obtained from the Army Corps of Engineers show iron oxide stains in cores from a depth of 820 feet (270 m). Apparently, fracture systems are extensive and very deep weathering occurs, at least in some parts of the batholith.

Several clues can be observed in the field to determine whether hard, unrippable rock is likely to lose strength rapidly and spall within a year or two. If the feldspars are clear, particularly at their boundry with adjacent mineral grains, rapid deterioration of the rock is unlikely. We found that staining rock with a 1 percent solution of p-aminophenol (Dodd 1955) was useful for indicating slight chemical weathering that might be undetected in a macroscopic field examination of rock (Hampton and others 1974a). In contrast, presence of iron oxide stains around biotite grains is of little value in predicting strength after exposure.

Increase in size and frequency of intergranular fractures is the single best expression of increasing physical weathering. For this reason, we would expect rock bulk density to be a good descriptor of degree of physical weathering (fig. 3). Unweathered rock, class 1, shows bulk densities in excess of 2.6 g/cc, essentially equal to the solid phase density of the individual minerals. Weathering to class 2 is a

chemical process involving initial hydrolysis of relatively easy-to-weather minerals--mainly biotite and plagioclase feldspars. These processes involve some expansion of mineral grains (Helley 1966; Clark 1967; Wahrhaftig 1965), which results in a lowering of bulk density to about 2.5 g/cc. Physical weathering processes exploit the small fractures resulting from the initial weathering, expand the fractures, and lower the bulk density to a range of 2.2 to 2.3 g/cc (weathering classes 3 and 4). At the same time, the rate of chemical weathering can increase due to the greater frequency of fractures. After this, the rocks disintegrate, forming grus. Little change occurs in bulk density until chemical weathering is quite advanced and secondary minerals make up a good proportion of the rock. Bulk densities of from 1.8 to 2.1 g/cc are common at this advanced stage of weathering (classes 6 and 7).

With increased physical weathering, we expected decreasing trends in rock hardness and strength along with the expected decrease in bulk density. These relationships were examined by plotting unconfined compressive strength of intact core samples and the hardness values developed by scleroscope against weathering classes. In addition, sonic wave velocities were measured on several core samples and related to weathering classes. These plotted relationships are presented in figures 4, 5, and 6. A more detailed study in Hampton and others (1978) essentially substantiates the results presented here.

The debates over insolation as an active process in physical weathering continue (Blackwelder 1933; Griggs 1936; Ollier 1969); however, both proponents and opponents of the insolation hypothesis agree that diurnal heating and cooling is most stressful to coarse-grained rocks, such as those in the Idaho batholith.

Exposed roadcuts are prime candidates for accelerated physical weathering resulting from diurnal fluctuations in temperature. This contention is supported by the observation that roadcuts produce substantial sediment during periods of no freezing or precipitation.

Nighttime relative humidities generally reach 100 percent, even during summer dry periods (data on file, Intermountain Station's research laboratory, Boise), and we might expect diurnal fluctuations in hydration of secondary minerals. These fluctuations may be a more important agent in producing sediment (grus) by granular disintegration than insolation.

Fire has also been proposed as a physical weathering phenomenon (Blackwelder 1927). This hypothesis has received recent support (Birkeland 1974).

We have observed flaking and granular disintegration of exposed boulders following wildfires and controlled burns in the Idaho batholith. These boulders may have been primed for such fire-induced weathering by previous chemical and physical weathering. Whether fresh (weathering class 1) rock would flake and disintegrate upon heating by fire is not known, although class 1 rock will readily flake upon heating with a Bunsen burner.

Chemical Weathering

The readjustment of bedrock to a surface environment involves chemical reactions brought about by the state of disequilibrium the rock is in when brought to the earth's surface by uplift and overburden removal. A few principal differences in surface environment dictate or define this state of disequilibrium: (1) lower pressure and temperature, (2) presence of copious amounts of water, (3) dissolved gases (principally O_2 and CO_2) in the water, and (4) presence of organic compounds, notably organic acids and chelating agents.

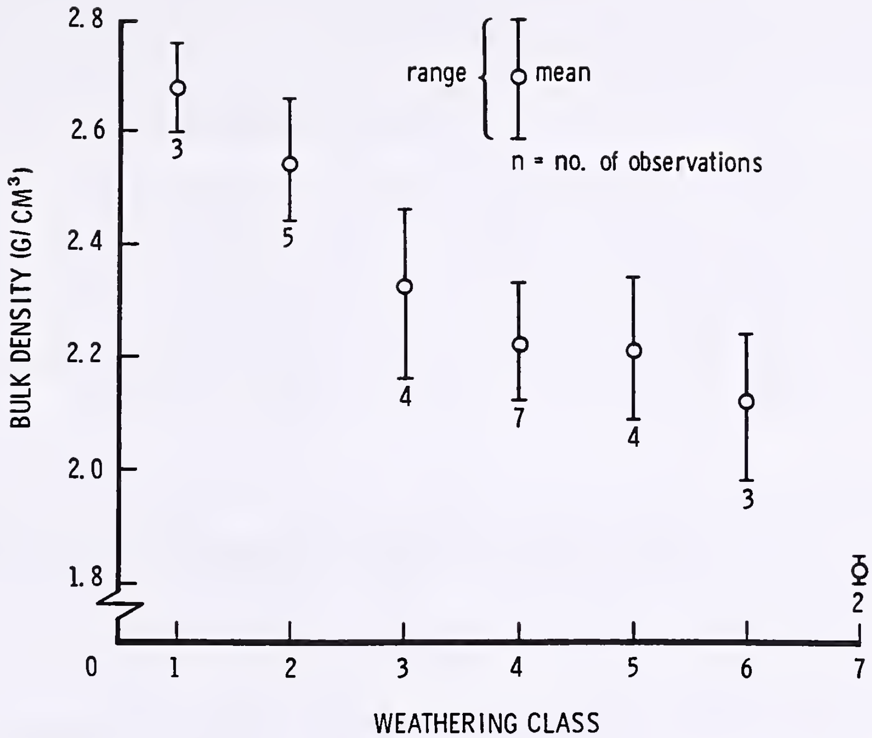

Figure 3.--Relationship between dry bulk density of bedrock (g/cc) and rock weathering class.

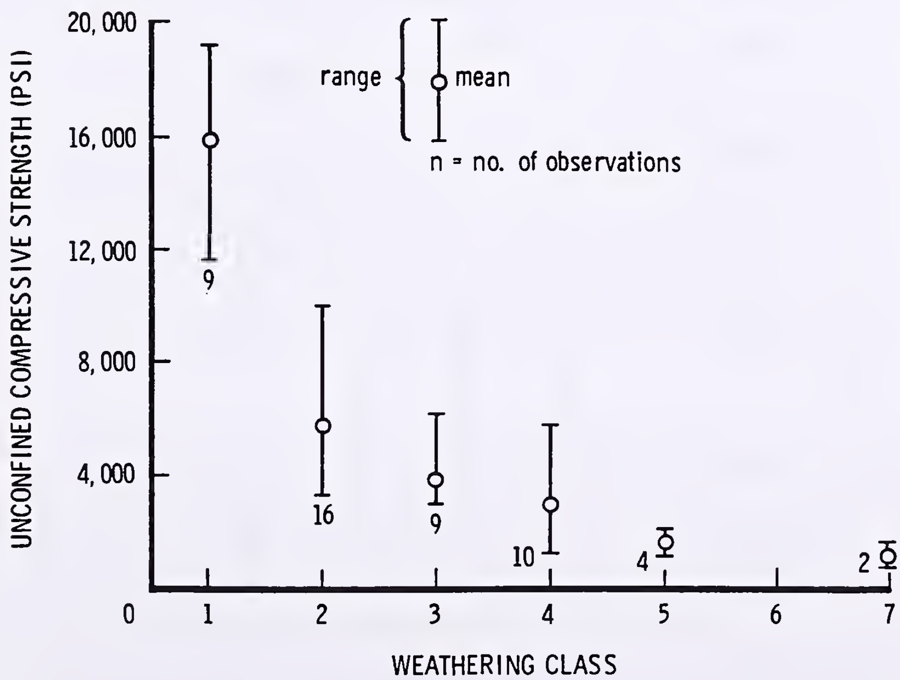

Figure 4.--Relationship between the unconfined compressive strength of rock cores measured in pounds per square inch (psi) and rock weathering class.

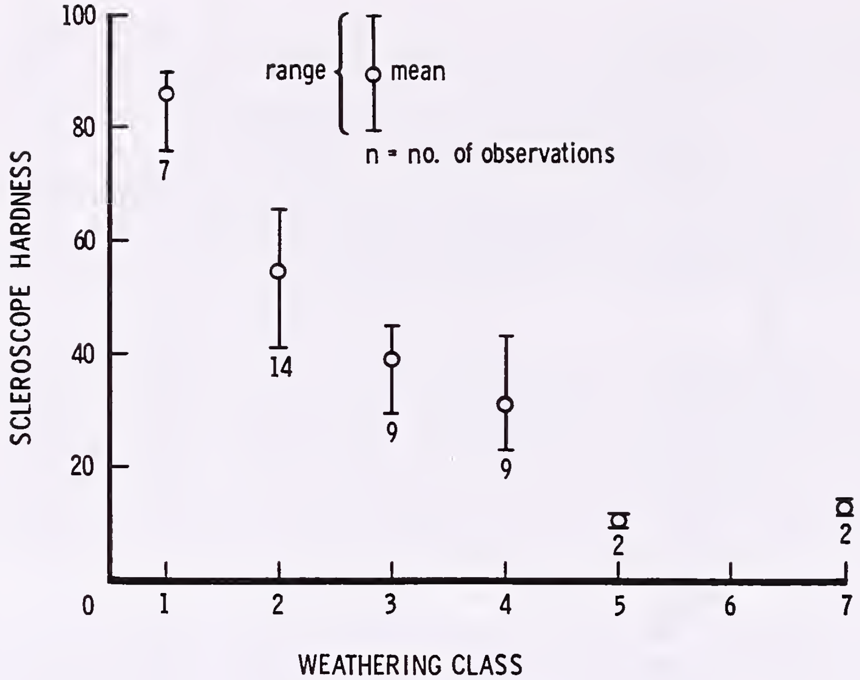

Figure 5.--Relationship between rock hardness measured with a scleroscope and rock weathering class.

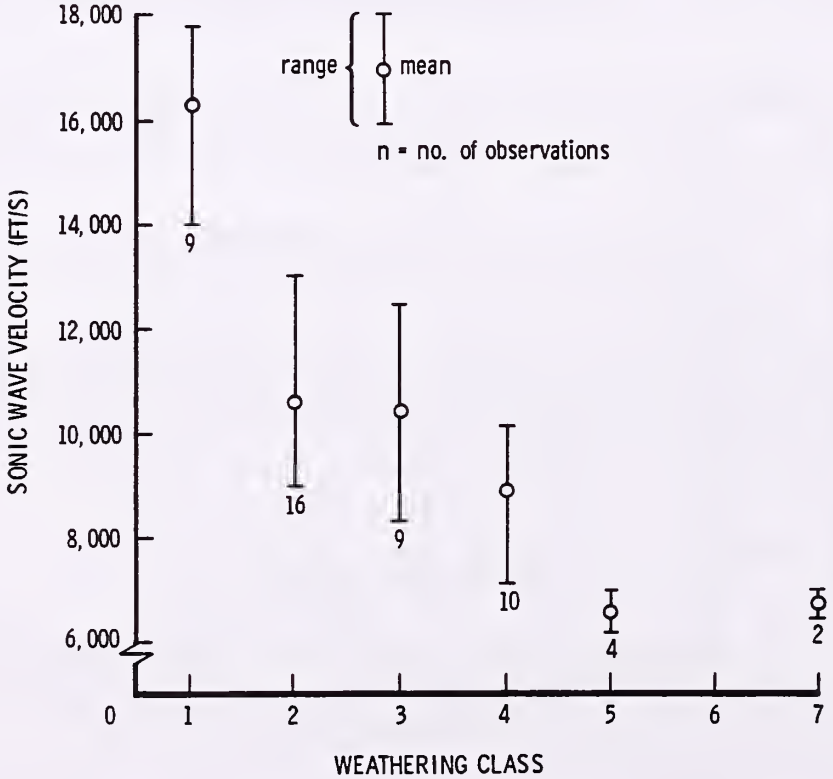

Figure 6.--Relationship between sonic wave velocity at an axial loading stress of 587 pounds per square inch and rock weathering class.

From theoretical considerations, the equilibrium byproducts of chemical weathering would be aluminosilicate clays, crystalline and amorphous oxides of silicon, aluminum, and base metals, especially calicium, sodium, iron, and potassium (Jackson 1965b). Studies of secondary minerals in soils in the Idaho batholith indicate the common solid by-products are kaolinite and halloysite, illite, allophane, and iron oxides. Primary quartz remains relatively unweathered.

The processes or reactions that produce those products may be thought of as desilication and alumination of primary minerals, since the reactions result in a net loss of silica relative to alumina (Jackson 1965a). Jackson (p. 16) identified three stages of desilication: "(a) mild desilication and alumination into phyllosilicate intergrades and allophane, (b) intermediate desilication (kaolinization), and (c) intensive desilication (laterization)." The products of mild and intermediate desilication in Idaho batholith soils are reflected in the clay mineralogy of batholith soils (Clayton 1974). The results of these soil processes will be discussed in a later section of this paper.

Weathering reactions and weathering products were studied below the soil in bedrock. This soil-bedrock boundary is easy to recognize in most instances, but is diffuse and poorly defined in weathering class 7 rock. The criterion for calling such material bedrock is the generally well-preserved fabric of interlocking mineral grains that give the appearance of granitic rock. This mineral grain accordance is not observed in soil peds in overlying C horizons, probably because of biotic and climatic (freeze-thaw) disturbances. Class 7 rock may also occur in gouge zones where less weathered adjacent rock is clearly below the lowest soil horizon. In other weathering classes, the soil-bedrock interface may be diffuse, but is readily recognized by the difference in structure, fabric, mechanical strength, and degree of oxidation.

A variety of techniques, including macroscopic (hand specimen) descriptions, thin section descriptions, chemical analysis, and x-ray analysis of pulverized rock samples were used to document chemical weathering. The important mineralogical differences between the weathering classes follow.

BIOTITE

Progressing from weathering class 1 to 6, biotite progresses from compact black grains to brown and golden grains having fringed edges. Biotites are absent in class 7 rock. No visible iron oxide stains emanate from biotites in class 1 rock. Iron oxide staining reaches a maximum in class 4 rock. In thin section, fresh biotites range in color from various browns to green and are pleochroic (fig. 7). They are strongly birefringent. Upon weathering, biotite grains have fringed edges, and iron oxide stains adjacent minerals (fig. 8). Upon further weathering, we have observed complete replacement of biotite by sericite (fig. 9) or possibly an iron-rich smectite-iddingsite (fig. 10).

This is in contrast to a study of ademellite weathering in the White Mountains, Calif., in which Marchand (1974) found little chemical weathering of biotite prior to grus formation. He attributes this to a drier climate, little evidence of chemical weathering, and a more pronounced physical weathering environment (Marchard, personal communication, 1976). In the Idaho batholith, we have never found grus formation without obvious chemical alteration.

Biotites in rock weathering classes 5 and 6 dye a violet to pink color when samples are treated with p-aminophenol. The purple-blue colors are characteristic of montmorillonoids and various shades of pink are characteristic of kaolin minerals, according to Dodd (1955). According to our interpretation, the presence of either color indicates argillation of the primary biotite without specifying the mineral because of the notable absence of smectites in batholith soils (Clayton 1974). However, this does not preclude the presence of smectites on weathered biotite grains.

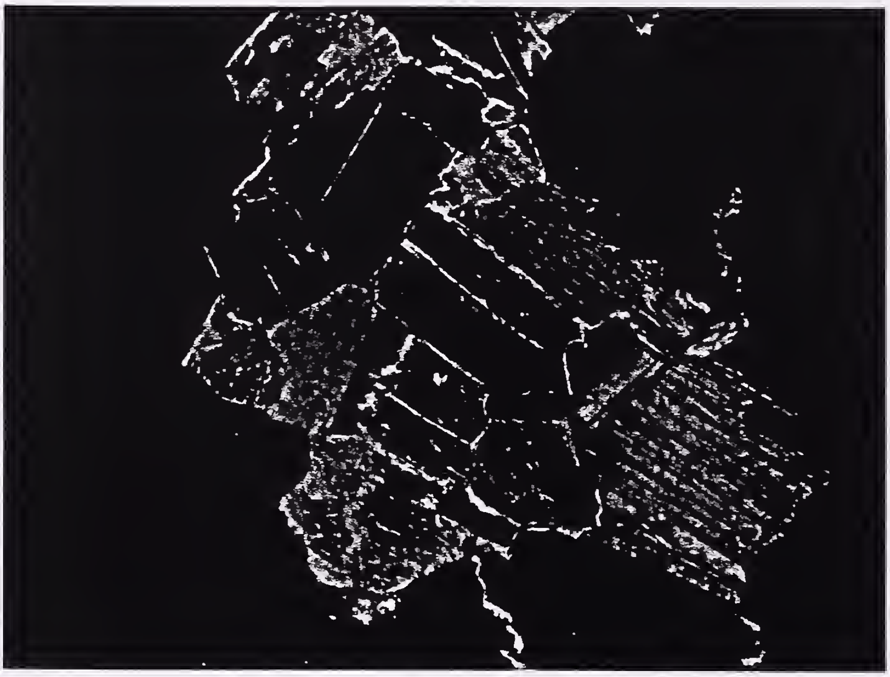

Figure 7.--Relatively fresh biotite grains exhibiting some weathering along grain boundaries (fringed edges). There is no apparent iron oxide staining of adjacent mineral grains. Magnification = 40X; nicols crossed. Transitional from weathering class 1 to 2.

Figure 8.--Weathered biotite grains exhibiting iron oxide staining of adjacent feldspars and partial alteration of biotite to iddingsite and smectite. Magnification = 40X; plane light.

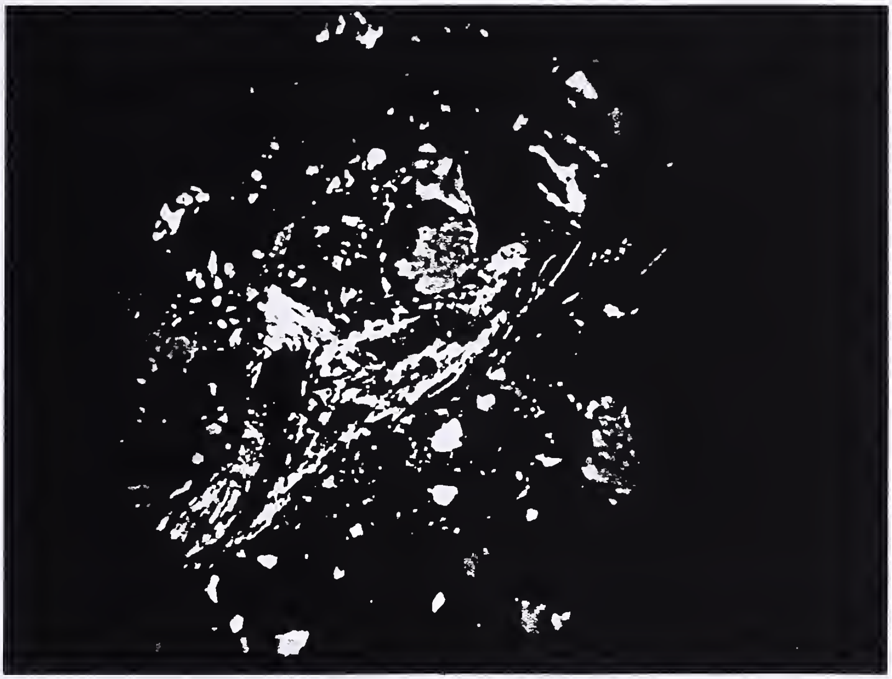

Figure 9.--Relatively complete sericite alteration of biotite in a rock weathering class 5. Magnification = 40X; nicols crossed.

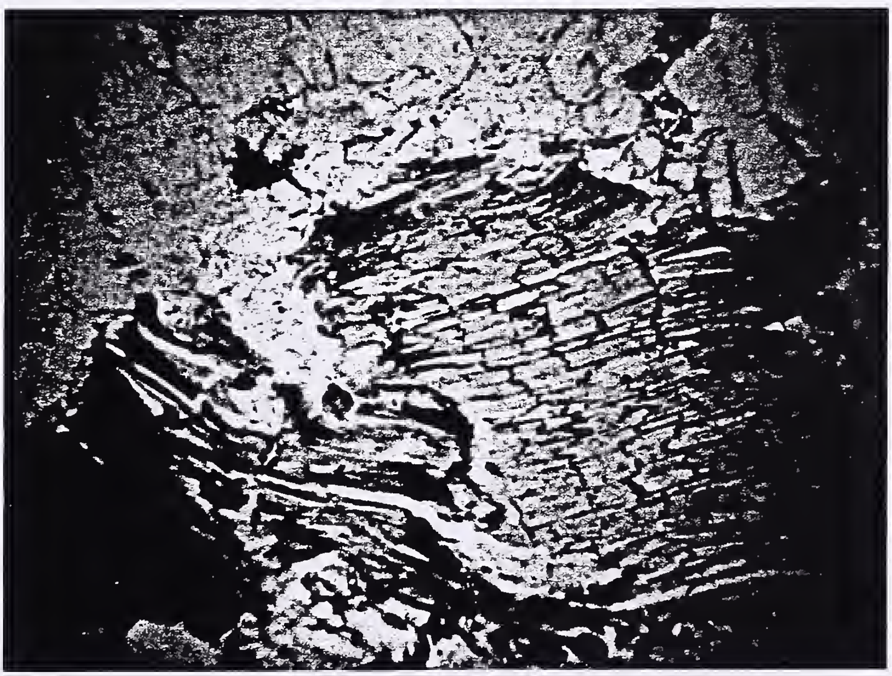

Figure 10.--Iron rich smectite-iddingsite alteration after biotite. Note oriented clay and iron oxide stains in microfractures of adjacent feldspar grain. Magnification = 100X; plane light. In a rock of weathering class 6.

In his summary of European and American research of biotite alteration in soil, Lucas (1962) described the principal transformations as either: (1) biotite to vermiculite and montmorillonite (both trioctahedral) by way of intermediate interstratified clays, or (2) biotite to chlorite, vermiculite and montmorillonite (all dioctahedral). Although all of these clay minerals have been reported in soils of the Idaho batholith (Clayton 1974), their presence is minor compared to kaolinite and illite. The dominance of kaolinite and illite would be expected on the basis of climate (Barshad 1966; Birkeland and Janda 1971; Clayton 1974).

FELDSPAR

It is often difficult to distinguish plagioclase feldspars from potassium feldspars in hand specimen. For this reason, the macroscopic descriptions of the feldspar weathering sequence combines all feldspars. Very fresh feldspars from weathering class 1 rock appear clear and transluscent. With increasing weathering (classes 2 through 4), feldspars become more opaque. Feldspars dye blue with p-aminophenol along their grain boundaries in weathering class 3 and darken to violet and pink as weathering increases. In weathering classes 5 and 6, the opaque appearance of the mineral grains changes to a powdery appearance and mechanical strength is greatly diminished.

One can work out the details of feldspar weathering by examining specimens in thin section. Most plagioclase crystals are in the range of Ab70 to Ab50, as determined by extinction angles of carsbad-albite twins (Kerr 1959). Plagioclase grains obviously are more argillized at lower weathering classes (3 and 4) than potassium feldspars. Kaolinite (or possibly halloysite) commonly forms pseudomorphs after plagioclase, distinguished by low relief and weak birefringence (fig. 11). Plagioclases in the Idaho batholith are commonly zoned and differential weathering of the more calcic core is frequently observed (fig. 12).

Sericitic mica (or perhaps illite) occurs as both an intergranular and a fine surficial replacement of plagioclase feldspar in thin section. The transition from plagioclase to illite is difficult to prove because sericitic mica is invariably present in the parent rock. Wilson (1975) points out that this transition is likely in confined saprolites because potassium is not removed by leaching.

Potassium feldspar argillation was not pronounced in weathering classes 1-3. In contrast, rather complete argillation of orthoclase was observed in higher weathering classes (fig. 13a, 13b, and 14). In places, orthoclase grains have extreme microbrecciation and partial argillation surrounded by a groundmass of primary quartz, orthoclase, and kaolinite. Large fractures are filled with chalcedony, suggesting hydrothermal water may have occupied the shear zone (fig. 15). Such extreme microbrecciation may result from strain associated with a localized shear zone, a common occurrence in the Idaho batholith.

Microcline is the common potassium feldspar in a few plutons within the Idaho batholith. Figures 16 and 17 show fresh and slightly weathered microclines from the Sawtooth Mountains, in the south-central part of the Idaho batholith. The minerals appear to be more susceptible to weathering along stress fractures.

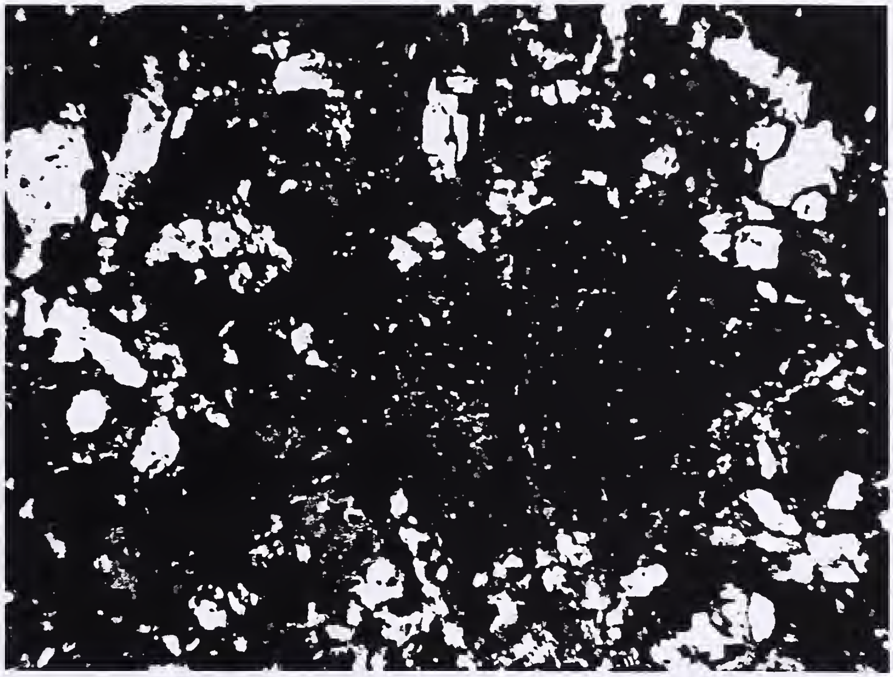

Figure 11.--Kaolinite pseudomorph after plagioclase showing relict twinning. Groundmass is highly fractured quartz, orthoclase, and altered biotite. Magnification = 40X; nicols crossed. Weathering class 5.

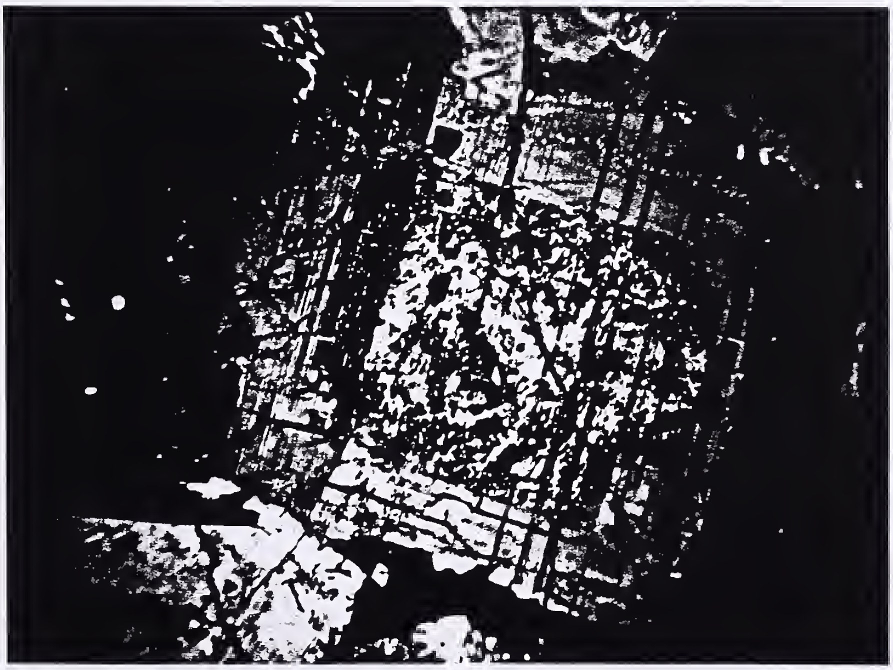

Figure 12.--Preferential weathering of the more calcic core of a plagioclase feldspar to kaolinite. Note relict twinning in the clay. Magnification = 40X; crossed nicols. Weathering class 4.

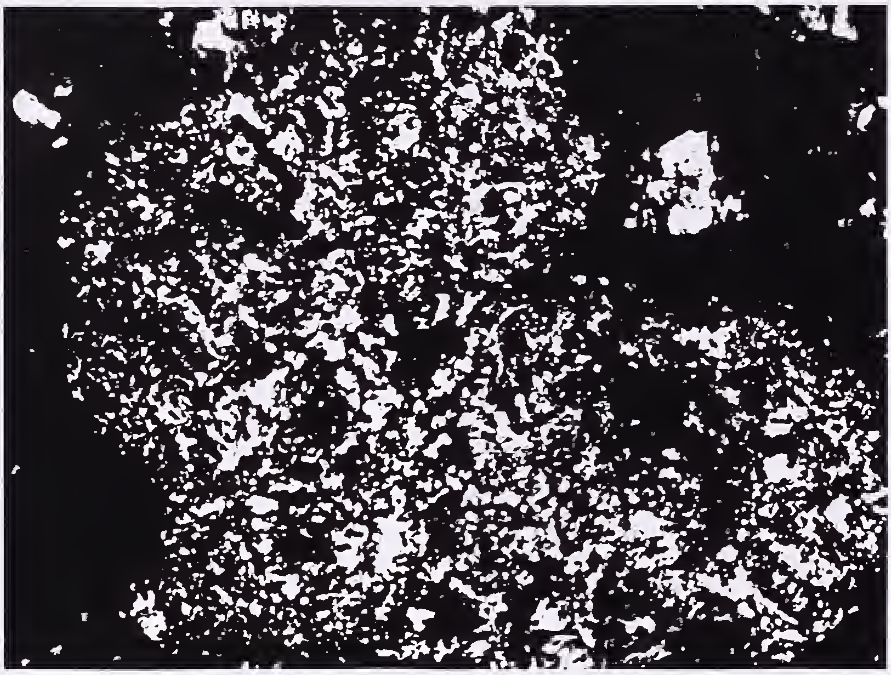

Figure 13a.--Complete argillation of a K-feldspar to kaolinite, an iron-rich smectite (?) and a venilet of iddingsite or iron oxide. Magnification = 100X; nicols crossed. Weathering class 6.

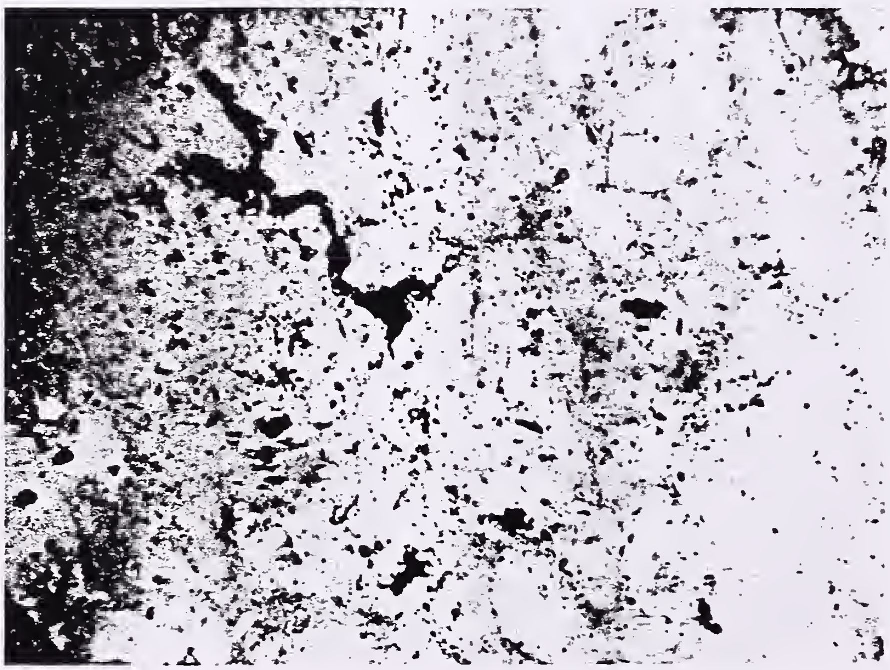

Figure 13b.--Same as figure 13a; plane light.

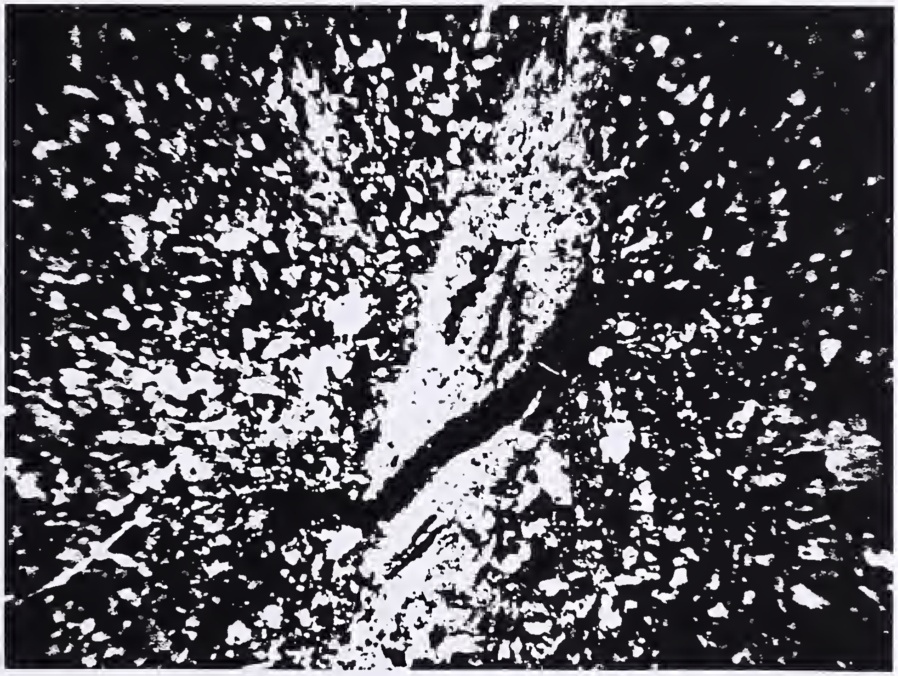

Figure 14.--A sericitized pseudomorph after orthoclase containing kaolinite, secondary mica, and iron oxide. Magnification = 40X; nicols crossed. Weathering class 5.

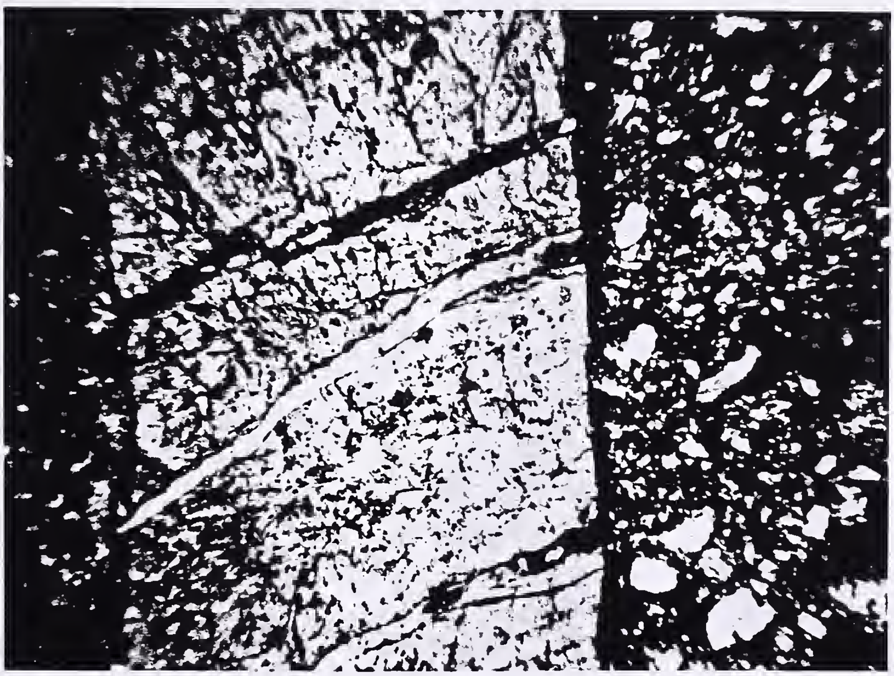

Figure 15.--A highly microbrecciated k-feldspar surrounded by a groundmass of primary quartz, orthoclase, and kaolinite. The large fractures are filled with chalcedony. Magnification = 40X; nicols crossed. Weathering class 5.

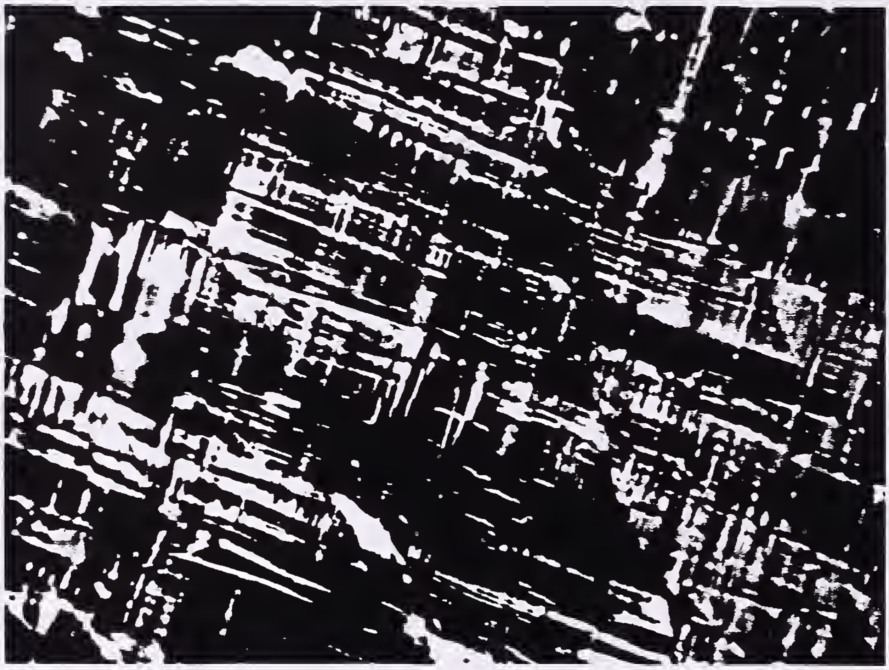

Figure 16.--Rather fresh microline exhibiting stress fracturing, but no chemical alteration. Magnification = 100X; nicols crossed. Weathering class 1.

Figure 17.--Microcline crystal showing incipient alteration along stress fractures. Magnification = 100X; crossed nicols. Weathering class 3.

QUARTZ

Alterations in quartz grains appear to be mainly mechanical, at least through weathering class 6. This may be due in part to the lack of obvious secondary mineral replacement of quartz. Highly fractured quartz grains are common in shear zones. Most quartz grains have an undulose extinction pattern, presumably strain induced during magmatic cooling (fig. 18).

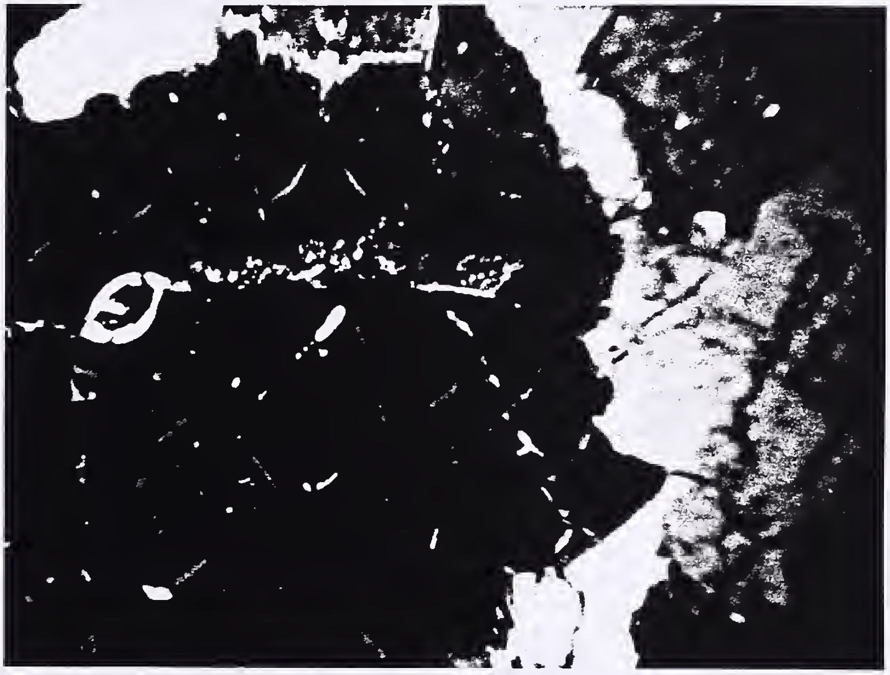

Figure 18.--Dark area at left is a large quartz grain near extinction (no transmission of polarized light). Light lines are strain induced and bound domains of crystal lattice deformation. This results in an undular extinction pattern when the microscope stage is rotated. Magnification = 100X; crossed nicols.

X-ray diffraction patterns from selected powdered rock samples show decreases in peak height and increases in peak diffuseness with increasing weathering (fig. 19). These changes are most apparent for plagioclase peaks from 3.16 to 3.26 Å (040 reflection), 3.66 Å ($13\bar{1}$), 3.78 Å (111) and at 4.04 Å ($20\bar{1}$). In general, the diffuseness (broad base, lack of symmetry) of the peaks is a better indicator of mineral weathering than peak height.

Slight changes in mineralogy of rocks of various weathering classes sometimes cause less distinct diffraction peaks, or the complete absence of a peak. For example, the 020 orthoclase reflection at 6.46 Å is absent in the class 1 sample (fig. 19), yet the $\bar{2}01$ reflection at 4.26 Å is present in all three samples.

Weathering class 1 rocks have a distinct 10 Å biotite peak. In contrast, class 7 rock has a 10 Å mica peak of equal intensity, but thin section observation indicates this to be entirely secondary mica, probably illite. Interestingly, rocks of intermediate weathering (classes 3 through 5) generally have less intense 10 Å peaks than either the fresh or highly weathered samples.

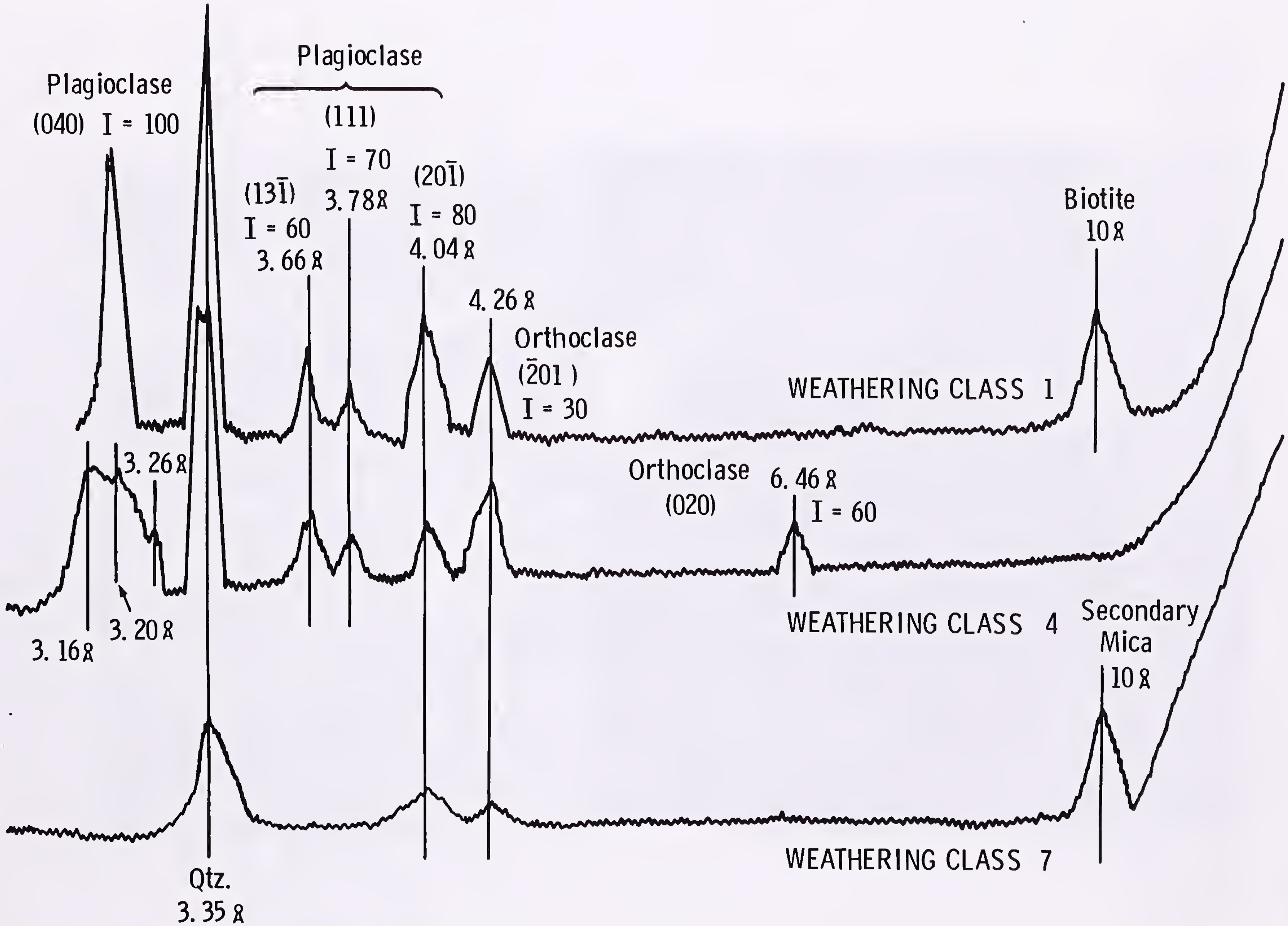

Figure 19.--X-ray diffractograms of powered bedrock showing differences in plagioclase feldspar peaks with increasing weathering. Note: d-spacing of crystal reflections are marked in angstroms. Miller indices of probable reflecting crystal faces are provided where appropriate. I-values are relative intensities; I-100 is the most intense peak.

Quartz peaks remain essentially unchanged through weathering class 5. There was some indication that the intensity of the quartz peak at 3.34 Å diminishes in weathering classes 6 and 7.

The only weathering products that appeared present in powdered rock diffraction patterns are 10 Å micas, although some mixed-layer clays are present in class 6 rock. We separated the clay fraction from soil and class 7 rock samples for x-ray examination, and class 7 rock are presented by Clayton (1974).

Weathering Progression the Idaho Batholith

Because of our studies of chemical, mineralogical, and physical changes brought about by physical and chemical weathering, we can propose typical weathering stages for batholithic rocks in Idaho. We feel that such a weathering progression would be similar for other batholiths with similar climates--winter freezing and summer dry periods. Such areas are extensive throughout the western United States and Canada.

Stage 1: Removal of overburden material.--Though not truly a weathering process, the rock expansion resulting from unloading associated with overburden removal forms small fissures large enough for water entry and subsequent weathering. This stage corresponds to weathering class 1. No chemical or mineralogic change has occurred in the rock, and bulk density is typically greater than 2.6 g/cc.

Stage 2: Initial chemical and physical weathering.--This stage is characterized by hydrolysis of iron and magnesium in biotites, oxidation of ferrous iron to ferric iron, loss of potassium, and a gain in water. Hydrolysis of calcic plagioclases is moderate, and initial hydrolysis of alkali feldspars is mild, but evident. Interstitial fracturing is evident in hand specimens, but rock is not spalling nor producing grus. Bulk density ranges from 2.3 to 2.5 g/cc. Stage 2 weathering corresponds to weathering classes 2 and 3.

Stage 3: Intense physical, moderate chemical weathering.--Following stage 2 weathering, the rock is "primed" for accelerated weathering because of the numerous pathways for water entry provided by stage 2 weathering. Physical weathering by cyclic freezing and thawing, and cyclic wetting and drying break down the weathered bedrock to grus. Processes of physical weathering have reached their apex in stage 3 weathering with appearance of thick spall rinds and ultimate grus formation. Spalling and formation of grus by granular disintegration requires that a rock surface be exposed to the atmosphere (Carroll 1970; Ollier 1969). Continued hydrolysis, oxidation and hydration of primary minerals, forms considerable quantities of clay minerals, allophane, and iron oxides. In hand specimen, feldspars appear powdery and stain blue and violet or pink with p-aminophenol. Clay pseudomorphs after feldspars, and micas and iddingsite alteration products, are commonly observed in thin section. Plagioclase feldspar x-ray peaks become diffuse and lose their intensity. Biotite peaks may disappear and secondary mica peaks show up in powdered rock x-rays. Stage 3 weathering involves the progression from weathering class 4 to weathering class 6, and bulk densities decrease ranging from 2.1 to 2.3 g/cc.

Stage 4: Intense chemical weathering.--Rocks that complete stage 3 weathering, but remain buried within or below a soil profile, are subject to intense *in situ* chemical weathering. In stage 4 weathering, the original rock fabric is preserved, but, with the exception of quartz, most other minerals have been altered to secondary weathering products. The result is a rock matrix of clays, crystalline and amorphous oxides of iron, silicon and aluminum, and embedded quartz grains. The rock is subject to plastic deformation when wet. Stage 4 weathering results in bedrock of weathering class 7. Bulk densitites of class 7 rock range from 1.8 to 2.1 g/cc.

HYDROTHERMAL ALTERATION OF BEDROCK AND SLOPE STABILITY

The geographical occurrence of bedrock of weathering class 7 was of considerable interest to us because of the strong correlation between class 7 bedrock and mass failures in the batholith (Megahan 1973; Day and Megahan 1976). Along the Middle Fork of the Payette River drainage in the southwestern Idaho batholith, 18 percent of 186 mass failures studied were associated with class 7 bedrock (unpublished data on file at the Intermountain Station's research laboratory, Boise). A generous estimate of the areal occurrence of class 7 rock in this drainage is 3 or 4 percent. In the Clearwater Forest in the northern Idaho batholith, Day and Megahan (1976) assigned 53 percent of 441 slides to a weathering class similar to class 7 of this study.

X-ray diffraction studies of the clay fraction from selected sites containing class 7 rock in the southern Idaho batholith show the expected clay mineral suites (kaolinite-halloysite and illite) in some cases and unexpected mineral assemblages (dominantly smectites and interstratified mixed-layer clays) in others. Clayton (1974) interpreted the smectite formation as a result of high silica potential and high base status associated with impeded drainage. Corroborative evidence of impeded drainage includes the common association of lepidocrocite with the smectite clays, and molar silica:alumina ratios twice the value found in freely drained soils with the kaolin-illite mineralogy.

Further field investigations revealed several sites fitting the field criteria for class 7 bedrock, but obviously occurring in gouge zones or localized shear zones. The clay alteration products often occur in seams within these zones, bounded by highly fractured (but chemically and mineralogically less altered) rock. These observations indicate that the clay is not a product of *in situ* weathering.

Boise State University, in cooperation with the Intermountain Station's research laboratory at Boise, investigated the mineralogical properties of several class 7 rock sites located in the southern Idaho batholith (Nichols and Nichols 1976)[4]. These sites included both shear zone locations with clay seams and locations where the argillized material appeared to result from in situ weathering. Sampling sites for these studies were concentrated in two areas, the Middle Fork of the Payette River in Valley County, Idaho, and the North Fork of the Boise River in Boise County, Idaho.

The Payette River sites are located along a major linear structural trend (Day and others 1974). Hot springs are located along this linear feature, often at the intersection of it and secondary linear features. According to Nichols and Nichols (1976), deep fracture zones associated with these lineaments provide the "significant plumbing systems for the deep convective circulation which is responsible for the hydrothermal alteration" (p. 13).

In contrast, the linear structural trends that control the course of the North Fork of the Boise River are more local in nature, and the width of the fractured rock and associated alteration is much less extensive.

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[4]Nichols, Clayton R., and Ann M. Nichols. 1976. Mineralogical investigation of hydrothermally altered plutonic rocks of the southwestern Idaho batholith. Final Rep., Coop. Agreement 12-11-204-113. Unpubl. Ms. on file, Intermt. For. and Range Exp. Stn., Boise, Idaho, 30 p.

The alteration products observed along the Payette River are dominantly kaolinite and illite with minor mixed-layer illite-montmorillonite. Individual grains of quartz, orthoclase, and plagioclase "float" in a groundmass of kaolinite and illite. Orthoclase crystals are clouded and sericitized along fracture faces. Plagioclases have undergone extensive internal alteration to kaolinite and sericite.

The alteration products observed along the Boise River differ markedly from those observed along the Payette River. The dominant alteration products at the Boise River sites are montmorillonite and clinoptilolite, a zeolite similar to heulandite. The zeolite occurs in veinlets or fracture fillings in areas of more intense alteration associated with extreme microbrecciation. Plagioclase feldspars exhibit zoned internal alteration to sericite (probably montmorillonite). Orthoclase crystals are fractured, but otherwise appear fresh. Biotites are generally fresh with minor alteration to montmorillonite at grain boundaries.

Nichols and Nichols (see footnote 4) consider the kaolinite-illite alteration along the Payette River to be more vigorous than the montmorillonite-clinoptilolite alteration along the Boise River. Their judgement is based upon the greater desilication and relative alumina enrichment and cation removal required for kaolin formation. In turn, they relate this more vigorous alteration to the more extensive fracture system associated with the large regional linear features along the Middle Fork of the Payette and the attendant hot water leaching.

Valley width is greater and the relief ratio is smaller along the Payette than along the Boise. Gentler sideslopes and deeper soils result, probably because of more extensive surface weathering. The resultant clay mineralogy is that expected from surface weathering and presumed to be in equilbrium with present-day pedogenic processes.

The above is not in disagreement with the hypothesis that the more extensive fracture system associated with a major lineament has resulted in more vigorous alteration. This does leave open the hypothesis that hydrothermal alteration in combination with subsequent surface weathering produced the kaolinite-illite suite.

The mass erosion problems associated with weathering class 7 bedrock appear to be related to poor soil drainage and loss of strength along the contact between class 7 rock and the adjacent less-altered rock. The high clay content of class 7 rock and lack of distinct jointing and fracturing result in a lower hydraulic conductivity than that in the overlying sandy soils and adjacent weathered bedrock. Most failures occur during spring snowmelt or long-duration rainstorms. The overlying soils become saturated after large volumes of water percolate to and are detained by the clay zone. Often the failure plane of the mass failure is at or directly adjacent to the clay zone.

Debris avalanches and rotational slumps are the common mass failures associated with class 7 rock, occurring in a 60:40 ratio, respectively (data on file at the Intermountain Station's research laboratory, at Boise). We feel that orientation of the clay seam or zone with respect to the slope is important. The hazard increases as dip angles increase and where the strike parallels the sideslope.

We have not observed inherent differences in slope stability associated with mineralogy. One might expect montmorillonitic clay zones to be less stable than kaolinitic zones because of the shrink-swell capabilities of montmortillonite. The more pervasive alteration of the kaolinite-illite clay zones along the Middle Fork of the Payette River may have masked any differences in erodibility associated with the clay mineralogy.

SOIL GENESIS AND MORPHOLOGY

Soils formed on slopes in the Idaho batholith are characteristically shallow to moderately deep, coarse textured, and either gravelly or stony. Typical soil profiles have A and oxidized C horizons underlain by weathered granitic bedrock. Coarse loamy sand and coarse sandy loam textures predominate. In residual soils, B horizons are uncommon on steep sideslopes. On gently sloping upland slopes that could be remnants of old erosion surfaces (see footnote 2), erosion rates are probably relatively low and cambic B horizons often are present. Argillic B horizons are weak and generally restricted to soils formed on alluvium or colluvium, and to slope gradients of less than 10 percent.

The common orders and suborders of soils that have been mapped in the Idaho batholith are (nomenclature follows Soil Survey Staff 1975):

Entisols

- Orthents
- Psamments
- Fluvents

Inceptisols

- Umbrepts
- Ochrepts
- Andepts

Alfisols

- Boralfs
- Xeralfs

Mollisols

- Borolls
- Xerolls

Important family differentiae in the Idaho batholith include the following classes:

(1) Particle size: sandy, sandy-skeletal, loamy-skeletal, coarse loamy, medial

(2) Mineralogy: mixed

(3) Temperature: frigid, mesic

(4) Depth: shallow

(5) Bedrock contact: lithic, paralithic

Of the state factors of soil formation presented by Jenny (1941), topography appears to most strongly influence soil morphology on slopes in the Idaho batholith. Erosion rates on the steeper slopes are sufficiently high that other state factors, such as climate, biota, and parent material, are not allowed maximal expression for pedogenic development over time. The topgraphic limitations result in "soils without normal profiles" (Jenny 1941). "This condition," he wrote, "is related to slope in such a manner that the greater the slope--as compared with the 'normal' slope--

the fewer the number of characteristic profile features and the more feeble their design. Soils of mountainous regions furnish the most conspicuous examples of this type."

We can better understand the reasons for these weakly developed soils by examining annual erosion and sedimentation rates from undisturbed, forested sites in the Idaho batholith. (Annual erosion and sedimentation rates expressed in cubic yards or tons per square mile per year have been adjusted to cm/1,000 years to clarify the following discussion.)

Megahan and Kidd (1972a) report annual erosion figures from erosion plot data ranging from 0.16 to 0.67 cm/1,000 years, for 2 years and 13 stations, in the west central batholith. We have erosion plot data (on file in the research laboratory at Boise) ranging from 0.02 to 5 cm/1,000 years, with a mean erosion loss of 0.86 cm/1,000 years. These figures are based on data collected over 4 years on 25 stations in the southwestern Idaho batholith. Much higher erosion has been observed (several cm per storm) during high intensity summer rainstorms or during rapid snowmelt (Megahan and Kidd 1972b).

In a study of sediment yields from small watersheds in the southern Idaho batholith, Megahan (1975) published figures on probable annual sedimentation rates indicating a 50 percent chance occurrence of 0.39 cm/1,000 years. A 10 percent chance occurrence event would yield a sedimentation rate of 1.5 cm/1,000 years. Sedimentation rates at these study sites, however, average less than one-half the erosion rates in the batholith. This discrepancy may be due in part to aggradation during valley building above the dams where the sediment is collected.

In any case, erosional losses exceeding 1 cm/1,000 years are to be expected in the batholith under present climatic stresses.

Marchand (1971) points out that major alluviations in valleys of the western cordillera (and hence erosional events on mountain slopes) occur during glacial rather than interglacial periods, even in unglaciated basins. Therefore, our short-term records of erosion rates are minimums when considered over a period of several tens of thousands of years.

These erosion rates take on considerable significance when time for pedogenic processes is considered. Birkeland (1974) indicated that the time required for steady state expression (maximal development) of A horizons ranges from 100 to 1,000 years, for cambic horizons several thousand to 100,000 years, and for argillic horizons, is greater than several hundred thousand years. Time for initial expression of these horizons would be one or two orders of magnitude less. It is readily apparent that the time required for even modal development of B horizons is simply not available on slopes losing 1 cm or more of soil each 1,000 years, when the total soil depth averages 50 to 60 cm.

Sampling and Laboratory Studies

Descriptive soil morphology and a variety of laboratory analyses were performed to meet the original study objectives of relating soil properties to bedrock weathering and fracturing characteristics and to geomorphological considerations. Sampling procedures assured variety in bedrock characteristics and geographic and climatic diversity. The necessity of sampling roadcut faces tended to bias sampling to ridgelines.

Several soil properties reflecting degree of soil profile development were examined. These properties were considered from the standpoint of topographic and climatic variables and slope stability (and thus age). Properties included: horizonation, color, amount of clay and clay mineralogy, free silica and alumina in the clay fraction, free iron oxides, water retention, percent organic matter, heavy mineral grain morphology, and pH. We were particularly interested in relating degree of soil development to weathering and fracture density characteristics of the underlying parent rock. Where topographic influences (steep slopes) are not the limiting factor in soil development, rates of bedrock weathering are limiting. These rates are also a function of climate and biota. Of secondary interest was the influence of topographic and climatic variables on soil genesis and bedrock weathering and alteration.

Soil Characteristics and Rock Weathering Relationships

The degree of horizonation and the distinctness of soil horizons are related to weathering classes. For example, two of three sites underlain by class 1 (hard, unweathered) rock have no soil formation except an accumulation of surface litter (01 horizon). The third site has a shallow A1 horizon and a transitional AC horizon containing 60 percent stone and rock. Bedrock at this site is highly fractured and the greater soil development (the presence of both an A1 and AC horizon) is likely attributable to the high density of fractures.

In contrast, soils overlying bedrock of weathering classes 4 through 6 generally have A11 and A12 horizons and two or more C horizons transitional to the bedrock. Soils usually exhibit paralithic contacts with bedrock.

The boundary between the C horizon and bedrock becomes less distinct as the weathering class of the underlying bedrock increases. Hence, soils overlying weathering class 1 or 2 bedrock have a much more distinct C-R horizon boundary than soils overlying bedrock of weathering class 5 or 6; however, as the weathering class of underlying bedrock increases, A-B, A-C, and B-C horizon boundaries often are clearer and more distinct.

Clarity and distinction of horizon boundaries is a function of time available for morphologic expression of pedogenic processes. Hence, we intuitively feel that soil formation has progressed over a longer period of time on well-weathered parent materials. Further, this explains why soils formed on more highly weathered bedrock have progressed further morphogically than soils of similar age on unweathered bedrock.

X-ray diffraction studies were carried out on the $<2\ \mu m$ fraction of all soils sampled and on selected bedrock samples ground to a fine powder. Clay mineralogy of the soils was found to be principally influenced by climate (and microclimate) (Clayton 1974). The 1:1 minerals (kaolinite and halloysite) predominate in well-drained soils of moderate or better development.

X-ray samples of bedrock indicate a variety of secondary minerals in weathering classes 6 and 7; however, only weathering class 7 samples gave sharp, distinct kaolin peaks. Sericitic mica (illite?), present as a fine surficial and intergranular alteration product of plagioclase feldspar and mixed-layer montmorillonite-illite are the common secondary silicates found in weathering class 6.

Although illite commonly appears in the clay fraction of soils (Clayton 1974) the mixed-layer clays are notably absent. We must presume that these mixed-layer clays are transformed in the soil, either to allophane, kaolin clays or to illite.

Free silica and alumina were determined in the clay fraction of soils. Stratifying the soils into dominantly kaolinitic clays versus dominantly montmorillonitic clays showed a marked difference in amounts of silica and alumina. Kaolinite clays average 1.6 times more silica and three times more alumina than montmorillonite clays (Clayton 1974); however, the molar silica/alumina ratio is 1.8 times greater in the montmorillonite clays relative to kaolinite clays. The larger ratio was explained by the greater solubility of silica relative to alumina and by the fact that in well-drained soils insufficient silica potential exists for montmorillonite formation.

We adjusted the percent of free silica and alumina in the clay fraction by bulk density and clay content of the soil to express the sum of the two in grams per cubic centimeter of soil. There is a generally increasing trend in free silica plus alumina per unit volume of soil, ranging from 0.002 g/cc in soils formed on weathering class 1 to 0.005 g/cc in soils formed on weathering class 7 rock. This increase is attributable to the slightly higher clay content of soils formed on rocks in higher weathering classes, particularly class 7.

Soil color and free iron oxide content of soils were analyzed to see if any relationships exist with degree of bedrock weathering. Zinke and Colwell (1963) found color to be the most obvious field characteristic indicative of the degree of weathering in upland soils in California. They found that color progresses from grayish brown to reddish brown with increasing soil development. This color progression is related to a progressive increase in free Fe_2O_3 content in the soil.

Birkeland and Shroba (1974) used the hue of Cox and B horizons to indicate intensity of oxidation (and hence age) in dating Quaternary soils. The hue progression is from 2.5Y or 5Y to 10YR to 7.5YR with time. This hue change is a progression from yellow to red and again reflects increasing iron oxide content in the C horizon.

In all of our soils the dominant hue in the oxidized C horizon is 10YR. Taking colors in weathered granitic rock is very subjective, however, because of the salt and peppery appearance due to the coarse grain size and differing colors of the constituent minerals. We did find a trend of increasing chroma from 1 or 2 to 4 with increasing weathering class of parent rock for soils of similar petrology. This relationship did not hold up universally in the soils we studied. As explained, free Fe_2O_3 content and iron oxide staining of bedrock appears to be related to the amount of biotite in the rock, not just to the degree of weathering.

Heavy mineral grain etching has been used to indicate degree of soil development or stage of weathering (Birkeland 1974). We separated heavy minerals in the fine sand and very fine sand fractions of selected soils for microscopic observation. Amphiboles and pyroxenes are almost completely absent in Idaho batholith rocks sampled; so we have no useful data on heavy mineral etching.

The pH of the soils studied shows a decreasing trend with increasing bedrock weathering class (fig. 20). We omitted A horizons from this analysis to diminish the confounding effect of organic matter. Decreasing pH with increasing weathering class can be expected due to the greater hydrolysis and subsequent leaching of released bases. Seventy percent of the variance in subhorizon pH can be explained by bedrock weathering class in this linear regression. The standard error of estimate is 0.29 pH units.

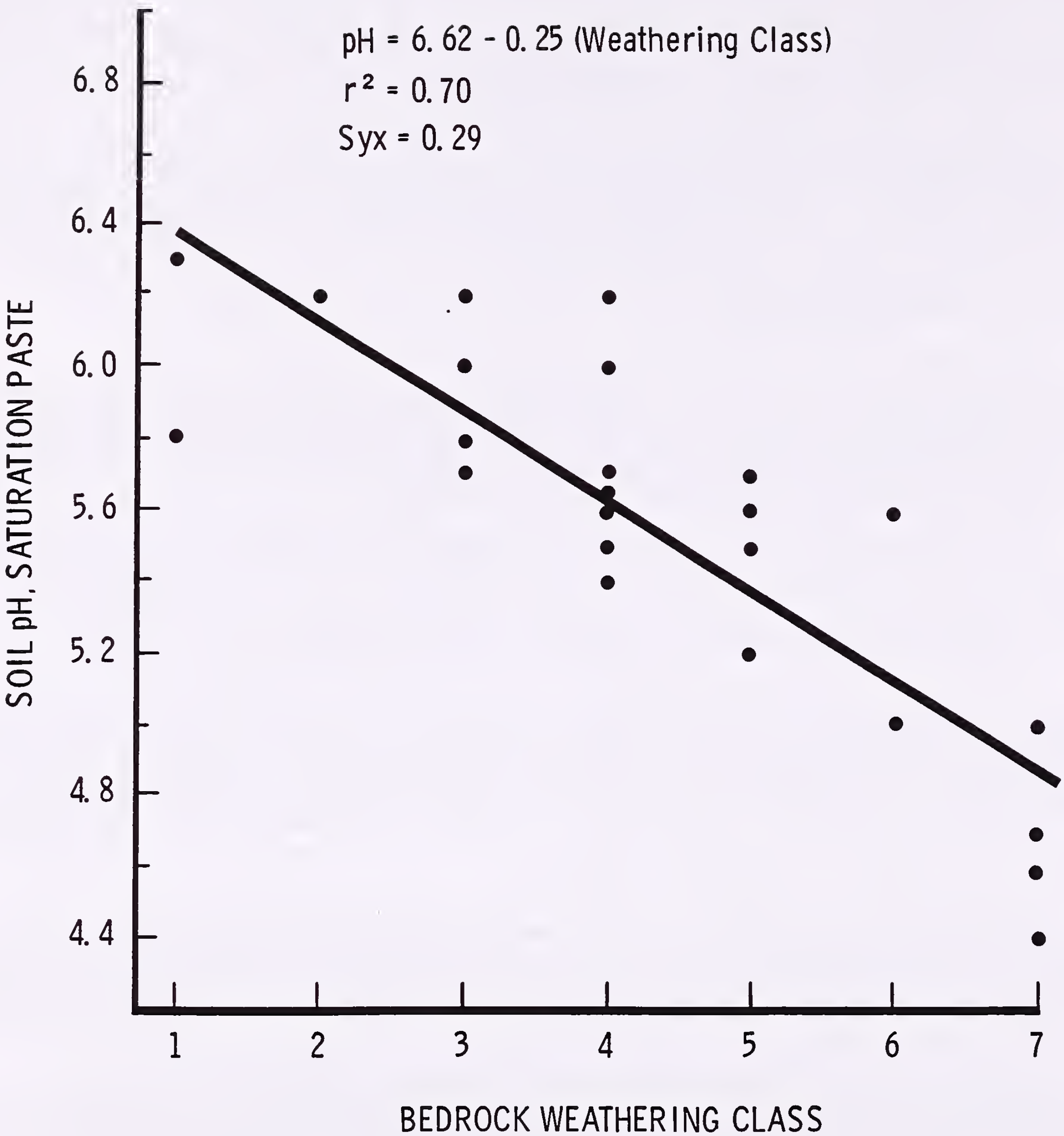

Figure 20.--pH of B or C horizons as a function of bedrock weathering class.

SUMMARY

The degree of weathering of batholithic rocks, as described by a previously devised seven-class system, appears to reasonably reflect physical and mineralogic changes in bedrock. We have described these changes as observed in hand specimen, thin section, and through chemical, physical, and x-ray analyses of fresh and weathered bedrock from the Idaho batholith.

After the unloading associated with overburden erosion, initial hydrolysis and oxidation of biotites provide sufficient pathways for water entry, a necessary pre-cursor to physical weathering. Rocks at or near the ground surface then progressively weather to grus, with physical weathering processes dominating chemical weathering. At depth, chemical weathering processes assume greater importance and the products of chemical weathering are better preserved. Biotites commonly weather to a degraded mica, then to a smectite-iddingsite product, and eventually to a 10 Å clay probably illite. Sericitic weathering products and, ultimately, kaolinization of feldspars are common.

Relationships between soil morphologic properties and bedrock weathering in the Idaho batholith are for the most part obscured by climatic and topographic influences, such as precipitation patterns, slope steepness, and internal soil drainage.

Slope steepness affects erosion rates and erosion strongly controls the time for pedogenic processes to differentiate soil horizons. Batholith soils are predominatly entisols, inceptisols, and weakly developed alfisols and mollisols, all reflecting lack of pedogenic development due to high erosion rates.

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Field Weathering Guide

The weathering classification of intrusive rocks published previously by Clayton and Arnold (1972) has been used extensively by land managers, primarily for relating bedrock weathering characteristics to slope stability. The research reported in this paper has provided for minor refinement and for expansion of this classification to include relationships between rock strength and weathering classes. In addition, the weathering stages published in this paper are related to weathering classes, providing a conceptual framework for the genesis of the various classes of weathered granitic bedrock.

Weathering classes are recognized in the field by the following criteria:

1. Change in rock color from that of the unweathered condition.

2. Mechanical strength of rock (how easily it can be broken manually)

3. Ease with which roots penetrate the rock matrix or fractures

4. How distinctly original jointing is preserved

5. The sound a rock hammer makes when striking the rock

6. Whether or not the rock is spalling (crumbling)

7. Whether or not the rock is plastic in nature when wet.

The above criteria should be used to identify the following classes:

Class 1, Unweathered Rock.--Unweathered rock will ring from a hammer blow; cannot be dug by the point of a rock hammer; joint sets are the only visible fractures; no iron stains emanate from biotites; joint sets are distinct and angular; biotites are black and compact; feldspars appear to be clear and fresh.

Class 2, Very Weakly Weathered Rock.--Very weakly weathered rock is similar to class 1, except for visible iron stains that emanate from biotites; biotites may also appear to be "expanded" when viewed through a hand lens; feldspars may show some opacity; joint sets are distinct and angular.

Class 3, Weakly Weathered Rock.--Weakly weathered rock gives a dull ring from a hammer blow; can be broken with moderate difficulty into "hand-sized" rocks by a hammer; feldspars are opaque and milky; no root penetration; joint sets are subangular.

Class 4, Moderately Weathered Rock.--Moderately weathered rock may be weakly spalling. Except for the spall rind, if present, rock cannot be broken by hand; no ring from hammer blow; feldspars are opaque and milky; biotites usually have a golden yellow sheen; joint sets are indistinct and rounded to subangular.

Class 5, Moderately Well Weathered Rock.--Moderately well weathered rock will break into small fragments or sheets under moderate pressure from bare hands; usually spalling; root penetration is limited to fractures, unlike class 6 rock where roots penetrate through the rock matrix; joint sets are weakly visible and rounded; feldspars are powdery; biotites have a light-golden sheen.

Class 6, Well-weathered Rock.--Well-weathered rock can be broken into sand-sized particles (grus); usually it is so weathered that it is difficult to determine whether or not the rock is spalling; roots can penetrate between grains; only major joints are preserved and filled with grus; feldspars are powdery; biotites may appear as thin silver or white flakes.

Class 7, Very Well Weathered Rock.--Very well weathered rock has feldspars that have weathered to clay minerals; rock is plastic when wet; no resistance to roots.

The four stages of weathering of granitic rock that are recognized in the Idaho batholith are based upon intensity and type of weathering (predominantly physical or predominantly

chemical). The following stages are related to weathering class and to decreases in bulk density and unconfined compressive strength as described below:

Stage 1: Removal of overburden material.--Though not truly a weathering process, the rock expansion resulting from unloading associated with overburden removal forms small fissures large enough for water entry and subsequent weathering. This stage corresponds to weathering class 1. There is no chemical or mineralogic change in the rock, and bulk density is typically greater than 2.6 g/cc; unconfined compressive strength is greater than 14,000 psi.

Stage 2: Initial chemical and physical weathering.--This stage is characterized by the hydrolysis of iron and magnesium in biotites, the oxidation of ferrous iron to ferric iron, a loss of potassium, and a gain in water. Hydrolysis of calcic plagioclases is moderate and initial hydrolysis of alkali feldspars is mild, but evident. Interstitial fracturing is evident in hand specimens, but rock is not spalling nor producing grus. Bulk density is in the range 2.3 to 2.5 g/cc; unconfined compressive strength ranges from 4,000 to 14,000 psi. Stage 2 weathering corresponds to weathering classes 2 and 3.

Stage 3: Intense physical, moderate chemical weathering.--Following stage 2 weathering, the rock is "primed" for accelerated weathering due to the numerous pathways for water entry provided by stage 2 weathering. Physical weathering by cyclic freezing and thawing, and cyclic wetting and drying break down the weathered bedrock to grus. Processes of physical weathering have reached their apex in stage 3 weathering with appearance of thick spall rinds and ultimate grus formation. Continued hydrolysis, oxidation, and hydration of primary minerals form considerable quantities of clay minerals, allophane, and iron oxides. Stage 3 weathering involves the progression from weathering class 4 to weathering class 6. Bulk densities decrease to the range 2.1 to 2.3 g/cc; unconfined compressive strength ranges from 1,000 to 4,000 psi.

Stage 4: Intense chemical weathering.--Rocks that complete stage 3 weathering but remain buried within or below a soil profile are subject to intense *in situ* chemical weathering. In stage 4 weathering, the original rock fabric is preserved, but, with the exception of quartz, most other minerals have been altered to secondary weathering products. The result is a rock matrix of clays, crystalline and amorphous oxides of iron, silicon and aluminum, and embedded quartz grains. The rock is subject to plastic deformation when wet. Stage 4 weathering results in bedrock of weathering class 7. Bulk densities of class 7 rock range from 1.8 to 2.1 g/cc; unconfined compressive strength is less than 2,000 psi.

Published as part of Research Paper INT-237, and as a refinement and expansion of the weathering classification in Clayton and Arnold (1972).

Clayton, James L., Walter F. Megahan, and Delon Hampton. 1979. Soil and bedrock properties: weathering and alteration products and processes in the Idaho batholith. USDA For. Serv. Res. Pap. INT-237, 35 p. Intermt. For. and Range Exp. Stn., Ogden, Utah 84401.

Weathering processes and products, hydrothermal alteration and soil genesis in the Idaho batholith are described. Stages of weathering are proposed and related to a previously published classification of rock weathering in the batholith. The relationships between weathering and hydrothermal alteration and hydrothermal alteration and various rock strength parameters are also described. Soil properities were not found to be strong related to bedrock weathering in the Idaho batholith.

KEYWORDS: Idaho batholith, rock weathering, hydrothermal alteration, soil formation, slope stability

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The Intermountain Station, headquartered in Ogden, Utah, is one of eight regional experiment stations charged with providing scientific knowledge to help resource managers meet human needs and protect forest and range ecosystems.

The Intermountain Station includes the States of Montana, Idaho, Utah, Nevada, and western Wyoming. About 250 million acres, or 90 percent, of the land area in the Station territory are classified as forest and rangeland. These lands include grasslands, deserts, shrublands, alpine areas, and well-stocked forests. They supply fiber for forest industries; minerals for energy and industrial development; and water for domestic and industrial consumption. They also provide recreation opportunities for millions of visitors each year.

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